HYDROLOGIC CONTROLS ON CARBON CYCLING IN ALASKAN COASTAL TEMPERATE RAINFOREST SOILS

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David V. D'Amore, B.A., M.Sc.

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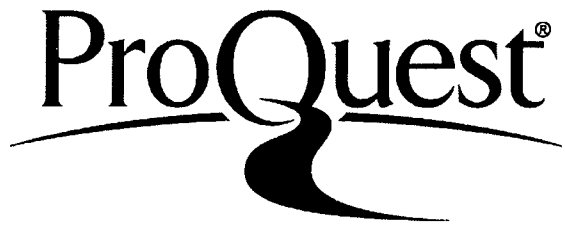

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By

David V. D'Amore

RECOMMENDED:

Advisory Committee Chair

Chair, Department of Forest Sciences

APPROVED:

Dean, School of Natural Resources and Agricultural Sciences

Dean of the Graduate School

August 22, 2011

Date

Abstract

The northern perhumid North American Pacific coastal temperate rainforest (NCTR) extends along the coastal margin of British Columbia and southeast Alaska and has some of the densest carbon stocks in the world. Northern temperate ecosystems such as the NCTR play an important role in the global balance of carbon flows between atmospheric and terrestrial pools. However, there is little information on key components of the forest carbon budget in this region. Specifically, the large pool of soluble carbon that is transferred from soils via streamwater as dissolved organic carbon (DOC) certainly plays a role in the total carbon balance in wet forests such as the NCTR. In order to address this information gap, I applied the concept of hydropedology to define functional landscape units based on soil type to quantify soil carbon fluxes and apply these estimates to a conceptual model for determining the carbon balance in three NCTR watersheds. The hydrologic gradient among ecosystem types served as a template for defining hydropedologic units for constructing a conceptual approach to constrain carbon budget estimates in entire watersheds. Replicated hydropedologic units were identified in three classes: sloping bogs, forested wetlands, and uplands. Estimates of annual soil respiration and DOC fluxes from the hydropedologic types were obtained through seasonal measurements combined with temperature-dependent models. Soil respiration fluxes varied significantly among the hydropedologic types, where soil respiration was 78, 178, and 235 g CO_2 m^{-2} y^{-1} in sloping bogs, forested wetlands, and uplands, respectively. Average DOC flux from sloping bogs, forested wetlands, and uplands was 7.7, 30.3, and 33.0 g C m^{-2} y^{-1}, respectively. These estimates of carbon export from the terrestrial ecosystem were combined with values of net primary productivity from remote sensing to calculate net ecosystem production (NEP). The average NEP estimated in three NCTR watersheds was 2.04 ± 0.81 Mg C ha^{-1}. Carbon loss as DOC was 10-30% of the total carbon flux from the watersheds confirming the importance of this vector of carbon loss in the NCTR. The watershed estimates indicate that forests of the NCTR serve as a carbon sinks consistent with the average worldwide rate of carbon sequestration in terrestrial ecosystems.

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Chapter 1: Introduction

1.1 Hydrologic controls of carbon fluxes in Alaskan coastal temperate rainforest soils

Forests contain large stocks of carbon and play an important role in regulating the global balance of carbon flows between atmospheric and terrestrial pools. A key question that guides the overall goal of the U.S. carbon cycle research program is to determine how large and variable the dynamic reservoirs and fluxes of carbon are within the Earth system, and how might carbon cycling change in the future. This goal has focused attention on the role of forests, and associated forest soils, in the storage and cycling of carbon. Soils store nearly five times the carbon contained in aboveground vegetation worldwide. However, the accurate accounting of the total store and the flow into and out of this enormous stockpile lags aboveground ecosystem component estimates.

The stock of carbon at northern latitudes is one of the larger pools of terrestrial carbon due to cold temperatures and impeded decomposition of organic material. The flow of carbon into northern temperate ecosystems has outpaced the decomposition of soil carbon in the recent past history of the earth. However, soil ecosystems are sensitive to temperature and the direction of the response for soil carbon stocks in the near future is highly uncertain given predicted changes in climate. Increased temperature and enhanced moisture conditions can lead to greater decomposition, which may outpace aboveground vegetation productivity. In this case, large carbon pools will be mobilized and exported to surface waters and the atmosphere providing a positive feedback to global atmospheric concentrations of carbon dioxide.

A great deal of emphasis has been placed on the boreal and Arctic regions due to 1) the storage of terrestrial carbon at these extreme northern latitudes and 2) predictions that climate warming will be greater at the higher latitudes. The coastal forests of the North American perhumid coastal temperate rainforest (NCTR) biome has received less attention due to the perceived small stock of carbon and the assumption of steady state carbon dynamics in this region. The NCTR is composed of humid coastal forests that are subject to abundant precipitation from prevailing Pacific atmospheric circulation patterns. The cool, wet conditions promote the development of dense coniferous forests, peatlands, and soils that have some of the densest carbon accumulations in the world.

In the NCTR, the major carbon flux into the terrestrial ecosystem is through the accumulation of biomass in large stature trees as well as the mosses that cover the extensive

peatlands. The quantity of carbon exported from the ecosystems of the NCTR has not been quantified, and it is assumed that the flux of carbon from the system through soil respiration comprises the bulk of the terrestrial output. However, the presence of darkly stained streamwaters draining the landscape indicates an export pathway for dissolved organic carbon (DOC) that must considered in the total carbon budget. In addition, the active glaciation, youthful landscape, and freshly exposed alpine areas create one of the largest exorheic areas for dissolved inorganic material in the world. Therefore, it is reasonable to assume that dissolved organic and inorganic carbon must be considered in determining the net ecosystem carbon balance in NCTR watersheds. Though often considered small fluxes in the overall ecosystem carbon balance, these fluxes may represent influential sources of carbon to aquatic systems in the NCTR. Heterogeneous mixtures of bedrock and till that are actively weathering along with exposed rock surfaces lead to inorganic carbon fluxes as bicarbonate. Dissolved CO_2 is also a potentially large carbon flux from terrestrial systems to proximal aquatic surface waters that may ultimately be respired in the aquatic system.

The research work undertaken in development of this dissertation was designed to address the goals of carbon cycle science in general and address some key aspects of the carbon cycle dynamics in the NCTR. The overall goal of the research project was:

> **To determine the cycles and fluxes of carbon from soils within key terrestrial ecosystem types in the North Pacific coastal temperate rainforest of southeastern Alaska.**

The experimental design and approach was crafted around the idea that dissolved forms of carbon export from soils were great enough in magnitude to impact the overall carbon budget of the NCTR. In order to frame this question appropriately, the flux of carbon through soil respiration was considered as a key component of the research design. The production and export of dissolved carbon is controlled by the cycles of soil organic matter and the interaction among soils, bedrock, soil organic matter, and soil respiration. The idea that the gaseous carbon flux as CO_2 varied inversely to dissolved carbon flux due to seasonal environmental effects such as temperature and precipitation on different soil types was the foundation for the hypothesis development and testing in the study.

The dissertation is arranged in five main chapters that describe the ecosystem, discrete ecosystem components, and present the results for several components of the carbon cycle. The dissertation culminates with a conceptual model that unites the components of carbon flux with net primary productivity to estimate the net ecosystem production across three research watersheds. Chapter 2 presents an overview of the components of the ecosystem that are used as end-member units to measure carbon flux and test the hypotheses in each subsequent chapter. The question "What is the relationship between terrestrial ecosystem type and carbon flux in the NCTR" can only be tested by defining ecosystem types that are readily delineated and scalable. The second chapter presents a literature review and overview of the experimental watersheds and sub-catchments from a perspective of how the large-scale hydrogeomorphic factors formed the watersheds, and the smaller scale influence of soils and vegetation patterns that dominate the sub-catchments. The second chapter established a justification for the use of the hydropedological type as representative for extrapolation of functional characteristics to similar scalable units that cover the NCTR.

Chapter 3 examines and quantifies soil respiration fluxes across the hydropedologic gradient. This chapter relies on the soil moisture influence among the soil types along with seasonal temperature patterns to build predictive models for soil respiration fluxes from soils. This chapter tests the idea that the soil respiration flux varies as a result of soil hydrologic conditions.

Chapter 4 examines how the composition of dissolved organic carbon in soil solution and stream water is influenced by fluctuations in soil moisture and temperature. The production and export of DOC is site specific and varies with temperature and soil physiographic setting. Predictions of DOC flux rely on physical indicators of water movement (i.e. slope) and precipitation. However, most fluxes of organic matter are linear in the measured ranges of discharge indicating that the soil organic matter storage acts as an infinite source over the observed climatic regimes. The soils and climate of the coastal temperate rainforest offer a very dynamic environment to evaluate the established relationship between soil organic carbon storage and stream water carbon flux. This environment offers the opportunity to evaluate the infinite source model of DOC production and export in soils.

Chapter 5 examines the patterns in the export of inorganic material and gases from the sub-catchment hydropedologic types and mainstem watersheds. Time series concentration measurements of dissolved elements were compared with atmospheric inputs to determine the net

retention or loss of elements in NCTR ecosystems. These measurements were also used to estimate the flux of inorganic carbon from the key end-member hydropedologic types of the watershed.

Chapter 6 extrapolates the functional response estimates among the component hydropedologic types in three watersheds to estimate total carbon flux. The model is based on the idea that watershed ecological processes occur in a predictable manner according to ecosystem type and can be modeled with measurements and estimates of key biogeochemical variables in the NCTR. Soil map units aggregated into categories that represent the catchment end-members were used to assign carbon flux estimates to discrete areas of the watershed. A conceptual carbon budget model was constructed from the total estimated carbon flux and an estimate of net primary productivity to provide a basis for determining net ecosystem production across the watersheds.

Chapter 2: Hydropedology of the North American coastal temperate rainforest[1]

2.1 Abstract

The North American perhumid coastal temperate rainforest (NCTR), which extends along the coastal margin of British Columbia and southeast Alaska, is characterized by intense orographic precipitation caused by Pacific storm systems striking coastal mountains. This intense precipitation influences the development of soil and vegetation communities, which in turn influence the transfer of terrestrial carbon and nitrogen (primarily in dissolved forms) and other nutrients to nearby aquatic ecosystems. These terrestrial subsidies of dissolved organic matter (DOM) are increasingly being recognized as an ecological function important to freshwater and marine ecosystems. The concentration and quality of DOM exported from NCTR soils varies according to soil end members (e.g. wetland vs. upland soils) that have formed along a topographic gradient from very flat organic soils to steep mineral soils. Hydropedology provides a template to evaluate the functions associated with water movement through soils along this gradient. We present a case study that demonstrates how hydropedology can be used to elucidate the physical mechanisms that influence DOM production and export in the NCTR. This information provides a foundation for modeling terrestrial DOM production and export under varying environmental conditions.

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[1] D'Amore DV, Fellman JB, Edwards RT, Hood E, Ping CL (in review) Hydropedology of the North American coastal temperate rainforest. In: Lin H (ed) Hydropedology: synergistic integration of pedology and hydrology.

2.2 The use of hydropedology in the North American coastal temperate biome

2.2.1 The North American Coastal temperate rainforest biome

The North American coastal temperate rainforest (CTR) biome extends along the coastal margin of the Pacific Northwest of the U.S.A, British Columbia and Alaska (Fig. 2.1). The perhumid northern portion of the CTR, the NCTR, is composed of the southeastern Alaskan panhandle and the north coast of British Columbia and represents the largest intact portion of the CTR. The NCTR is characterized by abundant precipitation from North Pacific storm systems that confront coastal mountains leading to intense orographic precipitation that influences terrestrial and aquatic ecosystems. Diverse terrestrial habitats and large fluxes of material to the coastal margin through plentiful surface water channels are defining attributes of the region. The NCTR has great ecological value for its abundant wildlife and diverse ecosystems that provide habitat for many endemic species of plants and animals (Cook et al., 2006). There is also an emerging appreciation for the transfer of dissolved organic matter (DOM) from terrestrial to freshwater and marine ecosystems along coastal margins such as the NCTR (Muller-Karger et al., 2005).

2.2.2 Applying hydropedologic techniques to establish a hydrologic observatory in the NCTR

The flow of water is the most fundamental component in the development and maintenance of ecosystem functions in the NCTR, yet the details of water storage, transport and delivery remain poorly understood at several spatial scales. This is despite the importance of the enormous flux of freshwater from terrestrial systems to the gulf of Alaska (Neal et al., 2010). Hydropedology is the multi-scale investigation of the source, storage, flowpath, residence time, availability, and spatio-temporal distribution of water in soils that occur in the Earth's critical zone (Lin, 2003). Hydropedology is also the study of how the transport of materials and energy by water can be quantified and understood (Lin et al., 2006). The application of hydropedology in the NCTR can increase our understanding of how the flow of material through terrestrial sub-systems is related to biogeochemical fluxes from entire watersheds. Establishing models of how terrestrial material influences downstream aquatic ecosystems is a priority for management of terrestrial and stream resources. Hydropedology can be used to create a conceptual framework to link individual soil types/terrestrial sub-systems with watershed biogeochemical function. In this

chapter, I present a framework based on the hydropedological research vision (Lin et al., 2006) and data from a hydrologic observatory network for designing and implementing hydropedological studies in the NCTR.

The NCTR is an excellent environment to develop a spatially distributed hydropedology model. There are two essential components of the hydropedological approach that can be applied to the NCTR. The first is landscape stratification using pedological mapping techniques. The second is the determination of hydrologic functions associated with these pedological units. Delimiting useful hydropedological units is a key to adequately attributing the influence of terrestrial hydrologic and biogeochemical functions on watershed outputs. With proper delineation, the variability in underlying biogeochemical processes within similar units can be minimized to capture the overall landscape response among units for development of robust watershed biogeochemical models. Testing models that describe terrestrial-aquatic biogeochemical linkages is very important given the increasing need to identify and predict the effects of changes in climate and land use on watershed nutrient cycles. The NCTR contains over 10,000 entire watersheds with a variety of different combinations of geology, weather, and vegetation, creating the need for adequate replicate systems for quantifying hydropedologic processes and testing resulting models on nutrient flux.

2.2.3 Integrating terrestrial and aquatic processes through hydropedology

This chapter details the implementation and use of a soil hydrologic observatory network in the NCTR that relies on the extensive soil ecosystem classification available through the Tongass National Forest (Tongass) soil resource inventory (USDA Forest Service, 1996). Our design for a hydrological observatory uses mapped geomorphic and vegetative patterns present on the landscape as a first approximation of functional units. We use these units to test how individual hydropedologic landscape units can be used to identify distinct functional responses in hydrodynamics and associated biogeochemical transport and transformation. We then examine whether these landscape units provide adequate resolution of cumulative biogeochemical and hydrological changes to extrapolate and aggregate these functions throughout an entire watershed. Finally, some tools for refining landscape stratification using topographic models for soil moisture are presented along with discussion of advantages and disadvantages of this approach.

2.3 Setting and details of the NCTR hydrologic observatory

2.3.1 Geology and climate

The accretion of terranes through the subduction of the Pacific plate under the North American continent combined with geologic faulting, deposition, and intrusion has formed a diverse assemblage of physiographic divisions along the extended island archipelago of the NCTR (Gehrels and Berg, 1994). Climate shifts and glaciation through several glacial epochs (Mann and Hamilton, 1995) left a pattern of mountain summits, fjordlands, wide glacial valleys, widespread deposits of glacial drift, and varying stream dissection along hillslopes. These post-glacial physiographic landscape features established the structural foundation for development of stream and vegetation patterns. The slope and parent material of many watersheds exerts a secondary influence on the trajectory of soil hydrologic patterns and associated development of soils and plant communities.

The rainforest climate of the NCTR is cool and wet across a wide range of latitude. Rainforests have at least 1400 mm of annual precipitation and cool annual temperatures (< 5 °C), (Alaback, 1996). The CTR contains four rainforest sub-zones: warm temperate, seasonal, perhumid, and sub-polar (Alaback, 1991; Fig. 2.1). The southern zones have fire as a source of stand replacement while fire is very infrequent in the northern zones. Annual precipitation increases and PET decreases from the south to north along the coast. As a result, moisture and temperature shift from limiting growth in the south to limiting decomposition in the north. Lower decomposition has resulted in enhanced organic matter accumulation at northern latitudes (Alaback and McClellan, 1992).

2.3.2 Vegetation and soils

Landform and climate are the principal components that determine soil geomorphic associations and plant distribution in the NCTR. Plant distribution patterns and ecology were strong influences in establishing landscape classification guides in the NCTR (Viereck et al., 1992; Meidinger and MacKinnon, 1989). The latest landscape stratification established subsection boundaries for the entire NCTR and neighboring parts of Canada using the National Hierarchical Framework of Ecological Units ('ECOMAP'; Cleland et al., 1997; Nowacki et al., 2001). The 'ECOMAP' delineation was guided by physiography, lithology, and surficial geology which are all factors related to the land's ability to process water (Nowacki et al., 2001).

Groundbreaking studies of landscape evolution and vegetation community development were undertaken in post-glacial landscapes of the NCTR (Chandler, 1942; Crocker and Major, 1955; Ugolini, 1968; Chapin et al., 1994). These early models of ecosystem development used the glacial chronosequence as a template for studying the dynamics of vegetation feedbacks to soils and the initiation of peatland as a climax state in the absence of disturbance due to the development of impermeable spodic horizons (Ugolini and Mann, 1979). Implicit in these models is the accumulation of water leading to water-logging of soils and development of deep organic accumulations. However, these models did not address seasonal cycles of water flow and accumulation in the soils and consequent changes in soils and vegetation that impacted the entire watershed.

The tree species of the region followed the development of soils with colonizing alders followed by conifers during the early part of the Holocene. Organic accumulations enabled the spread of Sitka spruce (*Picea sitchensis* (Bong.) Carr) and western hemlock (*Tsuga heterophylla* (Raf.) Sarg.) across the landscape from various glacial refugia. A climate shift to colder, wetter conditions approximately 4000 years ago promoted the development of extensive peatlands and the increase in red (*Thuja plicata* (Donn ex D.Don)) and yellow cedar (*Callitropsis nootkatensis* (d. Spach)) (Heusser, 1960). The current distribution of vegetation communities among different soil types is a response to fluctuating water tables that lead to well-drained or water-logged soils (Neiland, 1971).

Young, recently deglaciated areas quickly developed soil due to rapid organic matter accumulation and leaching. Spodosols can develop in approximately 200 years (Chandler, 1942) and are common on intermediate to well-drained soils. Histosols are also common and dominate areas of low slope that form the NCTR's abundant wetlands. The original studies established idealized principles for landscape evolution. However, there is much variability across the extended island archipelago. Soils of the region formed in the post-glacial Holocene deposits that are composed of glacial drift, colluvium and bedrock. The areas were periodically disturbed by windthrow which provides new parent material for soil formation (Bormann et al., 1995). Often modal concepts of the mapped soil series are not expressed throughout the mapping unit and a broad range of characteristics exists. Therefore, rather than relying on ideal expressions of soil type, I rely on general soil geomorphic relationships (i.e. higher order survey or taxonomic groupings) to identify soil types across broad landscapes.

2.3.3 A method to establish replicate hydropedologic units

Finely delineated terrestrial landscape attributes can be determined using high-resolution morphometric analyses such as LIDAR imagery. However, the delineation of these highly-resolved landscape attributes generally exceeds our ability to accurately measure the functional attributes of small features. The highest resolution of terrestrial ecosystem features is often desired, but cost, time, and technological limitations hamper efforts for precise delineation of landscape attributes and eventual mapping of discrete units in many cases. In the absence of fine-scale resolution of landscape features that can provide detailed soil and vegetation characteristics, we must derive soil-geomorphic relationships from existing data such as soil map unit classifications to establish initial hypotheses related to functional responses.

The highly-resolved landtype phase classification (Cleland, et al., 1997) which provides information on the distribution of fine-scale hydrologic classes, is not available in the NCTR. The best available information on soil hydrology is the drainage class assignment of soil map units based on soil morphology from the Tongass. The Tongass also has detailed resource inventories and spatially explicit maps for soil-vegetation assemblages associated with order 3 and 4 soil surveys. The availability of this inventory provides the building blocks for a preliminary hydropedologic framework across the region. The Tongass soil resource inventory provides a discrete approach (see Park and van de Geisen, 2004) to subdivide the coarse subsection structure (Nowacki et al., 2001; Fig. 2.2) into units that capture small-scale spatial variability within homogeneous geological formations. The discrete mapping approach recognizes the presence of a distinct spatial arrangement associated with pedologic and geomorphic structure that influences hydrologic gradients present on the landscape. There is a clear distinction in the hydrologic classes that reflects the strong influence of the distribution of water in soils on vegetation communities. We assume that hydrology exerts the primary control on biogeochemical transformations that vary across hydrologic classes, which seems reasonable given the abundance of research showing that hydrology exerts a first-order control on soil saturation (Stieglitz et al., 2003). The hydropedologic framework can be used to test the strength of association of hydrologic fluctuations and biogeochemical transformations with the provisional hydropedologic units. Plot based studies that quantify terrestrial-aquatic biogeochemical linkages within these hydropedologic units can be used to aggregate biogeochemical fluxes in individual units to entire watersheds through extrapolation using the Soil Resource Inventory.

The Tongass soil resource inventories were developed primarily through interpretations based on soil-vegetation relationships during soil and ecosystem mapping of the Tongass. The soil map units are closely associated with ecosystem types defined on landforms (USDA Forest Service, 1996) and are arrayed across parent material types such as bedrock, colluvium, and glacial till. There are many variants to established soil series due to the heterogeneity caused by small scale disturbance, such as windthrow and landslides. Soil map unit associations and complexes are also common because soil development can vary substantially due to the small-scale variation of soil forming factors caused by these localized disturbances. Wetlands are an obvious choice for a soil/vegetation class within the NCTR. Wetlands cover approximately 21% of the NCTR, though this estimate does not capture many of the wet forests that do not meet the jurisdictional requirements of wetland delineation. The readily available landscape classification of wetlands across the U.S. portion of the NCTR through the National Wetland Inventory (USFWS, 2009) makes applying this ecosystem classification quite useful. Within the available wetland classes, two are categorized as peatlands: sloping bogs, which are open areas with deep peat accumulations, and forested wetlands, which are occupied by overstory trees.

2.3.4 Establishing a hydropedologic observatory in the NCTR

We established a hydrologic observatory to implement a working model consistent with the research vision for hydropedology (Lin et al., 2006). We chose watersheds in three different ecological subsections as our core sites to address the higher order hydrogeomorphic control (Fig. 2.2). Peterson watershed, which is drained by Peterson Creek, is in the Stephens Passage glaciomarine terrace subsection and is composed primarily of slowly permeable glaciomarine sediments (Miller, 1973) along with bedrock outcrops that occur on moderate to low slopes (Fig. 2.2). Peterson represents an end-member type watershed dominated by wetlands (53% of watershed area). In contrast, McGinnis watershed, which is drained by McGinnis Creek, is primarily composed of recently deglaciated areas within the Boundary Ranges Icefield subsection and has low wetland coverage (<5% of watershed area). Eaglecrest watershed, which is drained by Fish Creek, is composed of intrusive volcanic and sedimentary rock in the Stephens Passage volcanic subsection and represents a mix of physiographic features from alpine to lowland wetlands.

Sub catchments were delineated within each watershed. These sub-catchments were uniform hydropedologic units defined through ecosystem classification and soil maps. We use the

term hydropedologic unit to define the combination of ecosystem classification with biogeochemical function. There are three potential hydropedologic units that can be identified through soil surveys and vegetation inventories. These hydropedologic units were delineated by soil and vegetation map units and verified in the field to assure adequate representation of end-members.

We assigned these classes as functional units by delineating three areas representing individual hydropedologic classes within each of the three watershed blocks (Table 2.1; Fig. 2.3). Therefore, similar hydropedologic units were located in each of the subsections. Histosols were identified in the sloping bogs and Histosols combined with Histic Cryaquepts were found in forested wetland sites. The hydrologic control section was defined by the presence and depth of the acrotelm boundary in the wetland soils (D'Amore et al., 2010). We chose uplands with soils located on relatively steep hillslopes defined by a lack of soil saturation with bedrock control as upland end-members.

Wetlands are roughly mapped by the National Wetland Inventory (NWI), but low resolution and inaccuracies dictate that field observations of soil patterns be used to more accurately delineate the boundaries of those classes in the field. The three hydropedologic units are closely associated with the soil map unit designation and associated drainage class (Table 2.1; Fig. 2.4).

We use the NWI nomenclature for classes of forested wetland and upland. We have established a "sloping bog" class as a modification to the NWI nomenclature that combines the features associated with fens and bogs. Sloping bogs have water flow derived from groundwater, similar to fens, which flows through vegetation communities more commonly associated with bogs. The slope bog is recognized in the Canadian wetland classification, but is not identified in the NWI. Therefore, we have established a class of hydropedologic unit that can be used across the entire NCTR.

The sub-catchments within each watershed were delineated to locate flow-gauging weirs to measure chemical export from each site (Fig. 2.3). The hydropedologic end-member units had distinct soil characteristics that were hypothesized to control soil hydrology and influence the overall hydrologic and biogeochemical behavior of the sub-catchment (Table 2.1).

2.4 A framework for integrated hydropedologic studies in the NCTR

Our integrated research fills a need to understand soil hydrology and watershed biogeochemical fluxes in the NCTR. Five key issues provided guidelines for describing our hydropedological research: structure, function, scale integration, and disturbance (Lin et al., 2006). We use these issues as a framework to present the information gathered through our research at the Juneau hydrologic observatory and evaluate the potential of future hydropedological research in the NCTR.

2.4.1 What are the hierarchical structures of soil and water in the NCTR?

The storage, flux, pathway, and residence time of water are distinctly different within our three classes, confirming that they span multiple dimensions of variability across the landscape. The NCTR has a moisture excess during most of the year, but the seasonal balance between precipitation and evapotranspiration creates consistent changes in seasonal water tables (Fig. 2.5). The water tables among the three groups were stratified as expected from the initial landscape partitioning (Fig. 2.6). Water tables increased from deep in the upland mineral soils to shallow in the forested wetland and sloping bog sites (Fig. 2.6). Although this relationship was expected, the consistency among the replicate units supported the premise that these units provide adequate first approximations of general water table fluctuations across the drainage gradient.

The water table depth in the mineral soils was limited by impermeable bedrock and slightly weathered regolith. Water moved through the soils on the 'upland' landscape positions, but there were fluctuations within the profile consistent with a groundwater flux in the lower horizons. The dominant flow was recharge and the movement of water downward through the spodic horizons. The movement of water was primarily through the matrix of mineral soil, with little or no overland flow. Even near-surface flow is limited and most of the water flow is directed laterally across the bedrock interface.

The flow regime of the forested wetland and sloping bog were regulated by the depth of the permeable, surface acrotelm horizon (D'Amore et al., 2010). There were two distinct cycles within these peatland soils; a distinct water table drawdown in June-August, and near surface saturation in the fall during higher rainfall (Fig. 2.7). The near surface saturation (<10 cm) was quite similar within the two wetland soils with nearly complete saturation of the soil profile, but forest wetland water tables dropped faster and to deeper depths during drawdown periods. The water table drawdown indicates the presence of an unknown water export pathway, either a much

deeper zone of flow within the forested wetland soils or evapotranspiration by trees. Rainfall events created sub-surface flow and pressure heads within the soil matrix that lead to ephemeral increases in water table height and incursions into the unsaturated soil zone (i.e. acrotelm) from the saturated zone below. These incursions are much more dramatic in the forested wetland due to the deeper arcotelm.

Biogeochemical transformations change with soil saturation. Understanding how soil saturation interacts with biogeochemical transformations within each hydropedologic unit provides insight into material transfers and biogeochemical transformations within a watershed. The duration and fluctuation of water tables in peatlands has been clearly linked to biogeochemical shifts (Strack et al., 2008). Although aerobic surface horizons persist in wetland soils even during periods of heavy rainfall (D'Amore et al., 2010), redox potential varied with soil saturation. The responsiveness of redox potential to changes in soil saturation demonstrates the highly dynamic shifts in microbial metabolism at the interface between the acrotelm and catotelm (Fig. 2.8). The anaerobic-aerobic boundary is a highly reactive zone for microbial activity, and hence, biogeochemical transformations (Gutknecht et al., 2006). The redox cycles vary seasonally and are influenced by soil type and fluctuations in soil saturation. Redox potential measurements are a surrogate measure for potential biogeochemical transformations within the soils and riparian sediments (Miller et al., 2006). For example, the soils are a rich source of reduced DOM, which is an important substrate for the microbial communities that oxidize and alter the nature of dissolved organic carbon (DOC) in soil and soil solution. The sequential processing of DOM by soil microbes contributes to electron transfers and is closely tied to the degradation of organic acids and other chemical transformations such as denitrification (Miller et al., 2009).

The depth and duration of aerobic conditions in the acrotelm is similar to other regions with extensive peatlands (Holden and Burt, 2003; Worrall et al., 2002, 2003), but the frequency and magnitude of water table fluctuations and resulting variability in redox potential are more dynamic in NCTR peatlands. Consequently, the detailed fluctuations in water tables provide a consistent pattern to identify discrete zones of biogeochemical activity.

2.4.2 How can function be integrated into the structure of the soil hydrologic system of the NCTR?

The NCTR provides an excellent natural laboratory for models such as the hydropedological model. It is the largest intact contiguous expanse of coastal temperate rainforest in the world, and contains thousands of watersheds largely devoid of major human disturbances. These watersheds contain a wide range of relative proportions of our defined units (e.g. 5-95% wetland) allowing us to verify watershed biogeochemical models across a broad geographic area. The application detailed here provides a first approximation for defining end-member hydropedologic units and using these units to describe biogeochemical cycles at the watershed scale. Thus, our examples provide a hierarchal structure to help explain biogeochemical cycling in individual landscape units with the ultimate goal of identifying biogeochemical signatures of an entire watershed.

Flowpath integration describes how the flowpath of water, and the material entrained in the flowpath, integrates stream and landscape heterogeneity (Fisher and Welter, 2005). Flowpath integration highlights how small-scale biogeochemical transformations that occur in three-dimensional space are integrated into net cumulative fluxes from the landscape (Fisher et al., 2004). The key concept in this approach is how nutrient retention and cycles associated with different ecosystem types control the composition of water flowing through the sequence of structural forms encountered on the landscape (Giblin et al., 1991; Fisher et al., 2007). Hydropedology and flowpath integration emerge from two different disciplines, but both strive to understand the linkage between the production, transformation and export of -DOM from terrestrial to aquatic ecosystems. Interdisciplinary concepts such as these are not often applied by a broad and diverse group of scientists, but remain within the discipline from which they have emerged. However, these concepts and the ability to integrate terrestrial and aquatic biogeochemical research are highly desirable in the NCTR. The concepts of terrestrial cycles and exports can provide a means to address the effects of alterations to functions in both systems and related maintenance of coastal marine food webs. Ultimately, the system must be viewed as a unified flowpath to achieve this goal (Fisher et al., 2004). The hydropedologic approach is a compromise that uses discrete terrestrial units to explain the continuous flow of material to aquatic ecosystems.

The hydropedologic landscape units are tools for organizing drainage units by function. The observed water table fluctuations and redox potential shifts within hydropedological units

can be linked with biogeochemical cycling and potentially to landscape unit function. The hydropedologic units are useful for describing DOM export from terrestrial to aquatic environments in the NCTR. In the same way, the hydropedologic unit approach can also provide insight into the quality of the DOM produced and exported from the landscape, which has important implications for productivity in downstream aquatic ecosystems.

The distinction between upland and wetland is a good predictive distinction for determining the DOM flux from a watershed (Mulholland and Kuenzler, 1979; Mulholland, 2003). Soil attributes, such as C:N have also been used effectively to predict watershed DOM export across a broad array of systems (Aitkenhead and McDowell, 2000). Our watershed stratification provides a good estimate for DOM concentrations (measured as DOC) using the wetland distinction. There is a strong relationship between wetland extent and DOC concentrations of in NCTR streams (D'Amore et al., in review; Fig. 2.9), which is apparent in the sub-catchment outlet streams for the hydropedologic units (Fig. 2.10). The hydrodynamics and intensity of the biogeochemical exchange in the acrotelm differs between the sloping bog and forested wetland sites and provides an explanation for the local strength of the association between depth to water table and soil DOC concentrations (D'Amore et al., 2010).

Integrating physical and biogeochemical cycles is essential in creating a useful tool to understand the effect of soil hydrologic cycles on DOM export. Surface organic horizons in both mineral and organic soils provide highly conductive flowpaths that facilitate the transfer of water and DOM through the soil and into surface water channels. Deeper, organic horizons have much lower conductivities and transfer water through the soil more slowly allowing for longer soil residence times for DOM and associated nutrients. The internal drainage of mineral soils shifts the chemical transformations from anaerobic to aerobic conditions, which is in direct contrast to peatland soils (Fig. 2.8). These conditions vary seasonally, as warmer periods and similar or reduced rainfall alter the late spring and early summer soil saturation dynamics.

Dissolved organic matter characterization provided a means for distinguishing biogeochemical signatures and potential ecological function among the hydropedologic units (Fellman et al., 2008; Fig. 2.11). For example, ordination analysis clearly distinguished the hydropedologic units into distinct classes based on the relationship between protein-like and humic-like fluorescence components (Fig. 2.11). These differences in the chemical properties of DOM were reflected in DOM bioavailability, as significant seasonal differences were observed both within and among hydropedologic units (Fig 2.12). Differences in dominant hydrologic

flowpaths and DOM lability suggest that these units have the potential to alter watershed-scale biogeochemical processes differently by influencing patterns in labile DOM delivery to streams.

Differences in chemical export within end-member hydropedological units and seasonal variability among these units has implications for the biogeochemical function of the entire watershed (Fellman et al., 2009b). For instance, low biotic demand and short soil residence time from predominantly surface soil flowpaths leads to a pulse of labile DOM during spring snowmelt for wetland sites, but less so for upland sites (Fig. 2.12). In the wetland sites especially, DOC concentrations can be quite low during the spring snowmelt because of simple dilution in relation to the mass of water. Competition between DOM transport and transformation is ultimately mediated by soil hydrologic flow rate, or contact time (Randerson et al., 2002). Thus, reduced DOM concentrations also may result from a stable water table that restricts organic matter mineralization and diminishes the pool of leachable DOM.

During the summer growing season, water table drawdown occurs in the bog and forested wetland and water moves predominantly through deep and less conductive flowpaths. High biotic demand for labile DOM by soil microbes and long soil residence time result in the delivery of mainly recalcitrant DOM to streams from wetland sites. Consequently, the quality and biodegradability of the streamwater DOM may vary dramatically with source material and season among the distinct landscape units (Fellman et al., 2009a). These observed differences in DOM biogeochemistry clearly demonstrate that these landscape units may have different functional roles in terms of biogeochemistry within the watershed.

Our work along the hydropedological gradient has also provided new insights into the role of discrete landscape units for providing labile organic matter to aquatic ecosystems. In particular, research in the NCTR has demonstrated the importance of wetlands for the export of labile DOM that can support stream heterotrophic productivity (Fellman et al., 2008; 2009a). This highlights a fundamental change in our view of wetland ecosystems, which have typically been seen as sources of highly degraded dissolved material. In addition to wetland ecosystems, new research has also shown that glacier ecosystems, which are common in the NCTR, can be important contributors of DOM to aquatic ecosystems in this region. The recognition of the importance of terrestrial nutrient subsidy to aquatic systems has been an important shift in our understanding of material flows in the NCTR. The transfer of material from the terrestrial to aquatic ecosystems in the NCTR provides important nutrient subsidies for both freshwater and marine ecosystems (Hood et al., 2009; Fellman et al., 2010). This finding also illustrates the

importance of understanding how hydropedological factors influence terrestrial biogeochemical cycles which can then be linked to soil water export pathways of dissolved reactive materials.

2.4.3 What is the potential for scaling functions across the complex landscape of the NCTR?

To constructively stratify the landscape, a broad grouping of 'subject' landscape types was chosen to represent geomorphic end-members. Finer distinctions could be made of these types, but because our intent was to upscale these fundamental units to the watershed scale, we used classes that could readily be predicted using available GIS data layers. The resolution of current digital elevation models (DEMs) and wetland and vegetation maps do not support a finer parsing of the landscape than the three fundamental types we used. These landscape units occur at different locations along the hydrologic gradient, and so represent rational ways to partition the landscape to study the functional linkages between soil process and hydrology.

The hydropedological stratification provides a framework for downscaling models of climate drivers to plots for modeling soil hydrology in connection with regional weather patterns. The parameter-elevation regressions on independent slopes model (PRISM, 2010) provides estimates of precipitation and temperature at a resolution of 0.8 km. Precipitation data obtained from PRISM models can be used to estimate evapotranspiration, which strongly influences the accumulation and flow of water in soils during the growing season. This information can provide support for developing watershed or regional water budgets by constraining estimates of fluctuations in soil water storage. The potential influence of deviations in soil saturation can also be applied to modeling plant responses across watersheds to guide vegetation-monitoring programs.

The highly dissected nature of the landscape of the NCTR is evident from coalescing flow on hillslopes during storms, and the integrated response of this water delivery system is quite profound. For instance, during intense storms, streamwater DOC concentrations often dramatically increase as terrestrial source pools and hydrologic flowpaths are different between baseflow and stormflow (Worrall et al., 2002; McGlynn and McDonnell, 2003). As soils become saturated, water moves through shallow soil layers or at the acrotelm/catotelm interface (Worrall et al., 2003) and differing flow paths can entrain DOM with different chemical properties (Fellman et al., 2009b). Thus, temporally transient water fluxes influence shifts in biogeochemical patterns. It is also possible that DOM export during storms accounts for a large fraction of the annual watershed flux, which highlights the importance of individual storms for

watershed biogeochemical budgets. Partitioning the watershed into hydropedological units and understanding their biogeochemical response to high flow events could be used to improve annual estimates of watershed-scale and regional DOM fluxes.

The distribution of frequently saturated soils can predict the presence of plant communities, which is especially important in the face of shifting populations of plants due to climate change (Hennon et al., 2006). Hydric soils have been successfully used as a predictor variable to describe differences in timber volume classes across the Tongass (Caouette and DeGaynor, 2005). However, a hydric soil class merely assigns a wet or dry class, but does not provide any details of how the underlying soil nutrient turnover may influence the specific vegetation assemblages. The hydropedological approach provides a framework to design experiments to examine small-scale processes in soils and the broader control on vegetation structure. This can then be linked to the biogeochemical properties such as DOM bioavailability to complete an assessment of the feedback mechanisms and interactions among all these key elements in watershed biogeochemical behavior. Yellow-cedar is a particularly good example of a potential application because its distribution is closely linked with soil moisture patterns. The species is not obligated to saturated soils (D'Amore and Hennon, 2006) but follows a niche strategy to persist in marginal conditions with low nutrient turnover and nearly saturated soil conditions (D'Amore et al., 2009). The prediction of future yellow cedar decline will rely on accurate spatial models of soil saturation as well as changes in those patterns over time.

2.4.4 How can information from diverse research studies be integrated into a hydropedologic model in the NCTR?

We provide a framework of hydropedologic units for predicting biogeochemical cycling that is similar to the geochemical catena (Johnson et al., 2000; Palmer et al., 2004). The influence of soil hydrology on biogeochemical cycles has not been clearly articulated in the coastal temperate rainforest research. Therefore, a working model using this approach and implementation of an integrated hydrologic observatory is useful for addressing goals associated with biogeochemical research in the NCTR. Our study outlines a case study to demonstrate how linking terrestrial and aquatic research can provide benefits to both fields. Traditionally, these disciplines have shared similar biogeochemical concepts such as the study of limiting nutrients, but measurements and perspectives regarding applications of these measurements vary (Grimm et al., 2003). One key example is the ability to derive information regarding wetland function from

structural aspects of wetland ecology (Bridgham et al., 1996). Aquatic ecologists have only recently explored the potential for wetlands to impact stream productivity beyond lakes and ponds.

The hydropedologic unit delineation can provide a link between disciplines as water flow is the key nexus between terrestrial and aquatic environments. Soil functional attributes can be applied to hydropedologic units for estimates of specific aspects of biogeochemical functions among individual hydropedologic units and these functions can be extrapolated to entire watersheds. For example, denitrification rates are influenced by soil factors such as texture and moisture content which control soil saturation and redox potential (Pinay et al., 2003). A hydropedological framework provides a means to estimate zones of high denitrification potential, such as riparian zones and seeps, based on theses soil factors. Using this approach, replicated studies can be designed to quantify denitrification rates that will enable more accurate estimates of nitrogen fluxes from these zones. The spatial resolution of these hotspots for denitrification can then be appropriately scaled to estimate the watershed nitrogen flux. The landscape units in our present model have been defined by generalized hydrologic fluctuations from the measurements of water table depths, but there are clear functional attributes associated with these three classes.

2.4.5 How does anthropogenic disturbance interact with hydropedologic functions?

The ecosystems of the NCTR are primarily driven by small-scale natural and anthropogenic disturbance over short time scales. Because fire is uncommon, wind is the primary driver of change in forest structure through disturbance (Kramer, 2001). The major anthropogenic disturbance in the NCTR has been timber harvest and to a lesser extent the construction of roads, especially through wetlands. In the southeast Alaskan portion of the NCTR, approximately 220,000 ha of forest have been logged. It is unclear how harvesting and road building has affected watershed hydrology and the delivery of water and materials to streams.

The extent of timber harvest covers approximately 3% of the total area of the Tongass and it may appear to not exert a large influence on the overall landscape. However, individual landscape features and soil types such as karst formations have been heavily used for timber harvest. The loss of large stature trees can alter evapotranspiration in stands and the subsequent timing and delivery of water from soils to streams. Although this type of change affects surface and subsurface water flow, its effects on soil attributes including soil development and biogeochemical cycling are unknown. The change in trajectory of soil development following

disturbance for individual landscape units is a key component of this question, and should be considered in overall hydropedology models.

Road networks associated with timber harvest are extensive in the NCTR. The impact of altered groundwater quality and flow and the alteration to material delivered to streams due to the presence of roads is not well understood. Therefore, road construction and the associated impacts to watershed hydrologic cycles remains a concern for forest managers. The limited number of existing hydrological studies on roads has been unsuccessful in identifying any major impacts to peatland (Kahklen and Moll, 1999) or mineral soil (McGee, 2000) hydrology. However, linear road features have potentially disrupted flowpaths and created biogeochemical interactions between soil water and crushed rock that may reduce the residence time of water within soils before delivery to the stream channel. Further, the increased rock-water interaction may alter the biogeochemical signature through oxidation of DOM and altered pH.

2.5 Research challenges and future applications

The most reliable DEM for the NCTR (southeast Alaska) is at a resolution of 60 m, which is too coarse for many continuous applications of catchment hydrologic and soil investigation used to delineate flowpaths. This level of detail provides some ability to establish coarse hydropedologic units based on interpreting zones of soil moisture and establishing hydrologic classes among soil map units. However, this topographic resolution does not provide an adequate level of detail to delineate flowpaths and intricate patterns of soil and water interaction necessary to understand biogeochemical cycling at small spatial scales. Consequently, there is a need for more detailed topographic information to better predict flowpaths of water and the hydropedology of landscapes in the NCTR.

A more refined model for soil geomorphic associations and linkage with landscape function could be derived through the application of a topographic wetness index guided by hydropedologic research. This approach can be used to differentiate wet areas from dry areas and delineate flowpaths that aggregate into definable units of similar hydrologic class (Fig. 2.13). The complexity of forested watershed hydrology is related to the structural diversity of the landscape. Landscape units that are operationally defined by quantitative geomorphometry are much more useful in delineating areas of water flow and coalescence. These models rely on accurate and preferably high-resolution digital elevation models along with good information on soil moisture patterns for calibration. Once more accurate topographic tools become available, we can link the

two approaches and create higher resolution maps and investigate smaller scale structural attributes of the soils.

2.6 References

Aitkenhead J, McDowell W (2000) Soil C:N ratio as a predictor of annual DOC flux at local and global scales. Glob Biogeochem Cycles 14:127-138.

Alaback PB (1991) Comparative ecology of temperate rainforests of the Americas along analogous climatic gradients. Revista Chilena Historia Naturel 64:399-412.

Alaback PB, McCellan MH (1992) Effects of global warming on managed coastal ecosystems of western North America. In: Mooney HA, Fuentes ER, Kronberg BI (eds) Earth system responses to global change: contrasts between North and South America. Academic Press, New York, pp 299-327.

Alaback PB (1996) Biodiversity patterns in relation to climate: the coastal temperate rainforests of North America Chapter 7 In: Lawford RG, Alaback PB, Fuentes E (eds) High-latitude rainforests and associated ecosystems of the west coast of the Americas. Ecological Studies 116, Springer, New York, pp 105-133.

Bormann BT, Spaltenstein H, McClellan MH, Ugolini FC, Cromack JK, Nay SM (1995) Rapid soil development after windthrow disturbance in pristine forests. J Ecol 83:747-756.

Bridgham SD, Pastor J, Janssens JA, Chapin CT, Malterer J (1996) Multiple limiting gradients in peatlands: a call for a new paradigm. Wetlands 16:45-65.

Caouette JP, DeGayner EJ (2005) Predictive mapping for tree sizes and densities in southeast Alaska. Landsc Urban Plan 72:49-63.

Chandler RF (1942) The time required for podzol formation as evidenced by the Mendenhall glacial deposits near Juneau, Alaska. Soil Sci Soc Am Proc 7:454-459.

Chapin FS, Walker LR, Fastie CL, Sharman LC (1994) Mechanisms of primary succession following deglaciation at Glacier Bay, Alaska. Ecol Monogr 64:149-175.

Cleland DT, Avers PE, McNab WH, Jensen ME, Bailey RG, King T, Russell WE (1997) National hierarchical framework of ecological units In: Boyce MS, Haney A (eds) Ecosystem management applications for sustainable forest and wildlife resources Yale University Press, New Haven CT, pp 181-200.

Cook JA, Dawson N, MacDonald SO (2006) Management of highly fragmented systems: the north temperate Alexander Archipelago. Biol Conserv 133:1-15.

Cowardin LM, Carter V, Golet FC, LaRoe ET (1979) Classification of wetlands and deepwater habitats of the United States. US Dept of Interior, Fish and Wildlife Service, Washington DC, 131 pp.

Crocker RL, Major J (1955) Soil development in relation to vegetation and surface age at Glacier Bay, Alaska. J Ecol 45:169-185.

D'Amore DV, Edwards RT, Biles FE (in review) Landscape controls on dissolved organic carbon concentrations of Alaskan coastal temperate rainforest streams. J Geophys Res

D'Amore DV, Fellman JB, Edwards RT, Hood E (2010) Controls on dissolved organic matter concentrations in soils and streams from a forested wetland and sloping bog in southeast Alaska. Ecohydrol 3:249-261.

D'Amore DV, Hennon PE, Schaberg PG, Hawley G (2009) The adaptation to exploit nitrate in surface soils predisposes yellow-cedar to climate change-induced decline and enhances the survival of redcedar. For Ecol Manage 258:2261-2268.

D'Amore DV, Hennon PE (2006) Evaluation of soil saturation, soil chemistry, and early spring soil and air temperatures as risk factors in yellow-cedar decline. Glob Change Biol 12: 524-545.

Fellman JB (2008) Dissolved organic matter in wetland soils and streams of southeast Alaska: sources, concentration and chemical quality. Dissertation, University of Alaska Fairbanks.

Fellman JB, D'Amore DV, Hood E, Boone RD (2008) Fluorescence characteristics and biodegradability of dissolved organic matter in forest and wetland soils from coastal temperate watersheds in southeast Alaska. Biogeochemistry 88:169-184.

Fellman JB, Hood E, D'Amore DV, Edwards RT (2009a) Seasonal changes in the chemical quality and biodegradability of dissolved organic matter exported from soils to streams in coastal temperate rainforest watersheds. Biogeochemistry 95:277-293.

Fellman JB, Hood E, Edwards RT, D'Amore DV (2009b) Changes in the concentration, biodegradability and fluorescent properties of dissolved organic matter during stormflows in coastal temperate watersheds. J Geophys Res 114:G01021, doi:10.1029/2008JG000790,2009.

Fellman JB, Spencer RGM, Hernes PJ, Edwards RT, D'Amore DV, Hood E (2010) The impact of glacier runoff on the biodegradability and biochemical composition of terrigenous dissolved organic matter in near-shore marine ecosystems. Mar Chem 121:112-122.

Fisher SG, Sponseller RA, Heffernan JB (2004) Horizons in stream biogeochemistry: flowpaths to progress. Ecology 85:2369-2379.

Fisher SG, Welter JR (2005) Flowpaths as integrators of heterogeneity in streams and landscape. In: Lovett GM, Jones CG, Turner MG, Weathers KC (eds) Ecosystem function in heterogeneous landscapes. Springer-Verlag, New York, pp 311-328.

Fisher SG, Heffernan JB, Sponseller RA, Welter JR (2007) Functional ecomorphology: feedbacks between form and function in fluvial landscape ecosystems. Geomorphology 89:84-96.

Frost P, Larson J, Johnston C, Young K, Maurice P, Lamberti G, Bridgham S. (2006) Landscape predictors of stream dissolved organic matter concentration and physicochemistry in a Lake Superior river watershed. Aquat Sci 68, 40-51.

Gehrels GE, Berg HC (1994) Geology of southeastern Alaska. In: Plafker G, Berg HC (eds) The geology of North America, Geological Society of America, Boulder, pp 451-467.

Giblin AE, Nadelhoffer KJ, Shaver GR, Laundre JA, McKerrow AJ (1991) Biogeochemical diversity along a riverside toposequence in Arctic Alaska. Ecol Monogr 61:415-435.

Grimm NB, Gergel SE, McDowell WH, Boyer EW, Dent CL, Groffman P, Hart SC, Harvey J, Johnson C, Mayorga E, McClain ME, Pinay G (2003) Merging aquatic and terrestrial perspectives of nutrient biogeochemistry. Oecologia 442:485-501.

Gutknecht JLM, Goodman RM, Balser TC (2006) Linking soil process and microbial ecology in freshwater wetland ecosystems. Plant Soil 289:17-34.

Hennon P, D'Amore D, Wittwer D, Johnson A, Schaberg P, Hawley G, Beier C, Sink S, Juday G (2006) Climate warming, reduced snow, and freezing injury could explain the demise of yellow-cedar in southeast Alaska, USA. World Resour Rev 18:427-450.

Heusser CJ (1960) Late-Pleistocene environments of North Pacific North America. American Geographic Society Special Publication 35.

Holden J, Burt TP (2003) Hydrological studies on blanket peat: the significance of the acrotelm-catotelm model. J Ecol 91:86-102.

Hood E, Fellman J, Spencer RGM, Hernes PJ, Edwards R, D'Amore D, Scott D (2009) Glaciers as a source of ancient and labile organic matter to the marine environment. Nature 24:1044-1048.

Johnson CE, Ruiz-Mendez JJ, Lawrence GB (2000) Forest soil chemistry and terrain attributes in a Catskills watershed. Soil Sci Soc Amer J 64:1804-1814.

Johnston C, Shmagin, B, Frost P, Cherrier C, Larson J, Lamberti G, Bridgham S (2008) Wetland types and wetland maps differ in ability to predict dissolved organic carbon concentrations in streams. Sci Total Environ 404:326-334.

Kahklen K, Moll J (1999) Measuring effects of roads on groundwater: five case studies. USDA Forest Service, Technology and Development Program, SDTDC 9977-1801, San Dimas, CA.

Kramer MG (2001) Maritime windstorm influence on soil process in a temperate rainforest. Dissertation, Oregon State University.

Lin H (2003) Hydropedology: bridging disciplines, scales, and data. Vadose Zone J 2:1-11.

Lin H, Bouma J, Pachepsky Y, Western A, Thompson J, van Genuchten R, Vogel H J, Lilly A (2006) Hydropedology: synergistic integration of pedology and hydrology. Water Resour Res 42:W05301, doi:10.1029/2005WR004085.

Mann DH, Hamilton TD (1995) Late Pleistocene and Holocene paleoenvironments of the North Pacific Coast. Quat Sci Rev 14:449-471.

McGee KL (2000) Effects of forest roads on surface and subsurface flow in southeast Alaska. Thesis, Oregon State University.

McGlynn BL, McDonnell JJ (2003) Quantifying the relative contributions of riparian and hillslope zones to catchment runoff. Water Resour Res 39:1310, doi:10.1029/2003WR002091.

Meidinger D, MacKinnon A (1989) Biogeoclimatic classification-the system and its application. In: Ferguson DE, Morgan P, Johnson FD (eds) Proceedings-land classifications based on vegetation: applications for resource management. USDA Gen Tech Rep INT-257, Moscow ID, pp 215-222.

Miller MP, McKnight DM, Cory RM, Williams MW, Runkel RL (2006) Hyporheic exchange and fulvic acid redox reactions in an alpine stream/wetland ecosystem, Colorado Front Range. Environ Sci Tech 40:5943-5949.

Miller MP, McKnight DM, Chapra SC, Williams MW (2009) A model of degradation and production of three pools of dissolved organic matter in an alpine lake. Limnol Oceanogr 54:2213-2227.

Miller RD (1973) Gastineau Channel formation: a composite glaciomarine deposit near Juneau, Alaska. Geological Survey Bulletin, Contributions to stratigraphy, USGS Report #1394, Washington DC.

Muller-Karger FE, Varela R, Thunell R, Luerssen R, Hu C, Walsh JJ (2005) The importance of continental margins in the global carbon cycle. Geophys Res Lett 32:L01602, doi:10.1029/2004GL021346.

Mulholland P, Kuenzler E (1979) Organic carbon export from upland and forested wetland watersheds. Limnol Oceanogr 24:960-966.

Mulholland PJ (2003) Large-scale patterns in DOC concentration, flux, and sources. In: Findlay S, Sinsabaugh R (eds) Aquatic Ecosystems: interactivity of dissolved organic matter, Elsevier Science, San Diego, pp 139-159.

Neal EG, Hood E, Smikrud K (2010) Contribution of glacier runoff to freshwater discharge into the Gulf of Alaska. Geophys Res Lett 37:L0604. doi:10.1029/2010GL042385.

Neiland BJ (1971) The forest-bog complex in Southeast Alaska. Vegetatio 22:1-63.

Nowacki G, Shephard M, Krosse P, Pawuk W, Fisher G, Baichtal J, Brew D, Kissinger E, Brock T (2001) Ecological subsections of southeast Alaska and neighboring areas of Canada. Anchorage AK, US Department of Agriculture Forest Service, Alaska Region 10, R10-TP-75, 306 pp.

Palmer SM, Driscoll CT, Johnson CE (2004) Long-term trends in soil solution and stream water chemistry at the Hubbard Brook Experimental Forest: relationship with landscape position. Biogeochemistry 68:51-70.

Park SJ, van de Giesen N (2004) Soil-landscape delineation to define spatial sampling domains for hillslope hydrology. J Hydrol 295:28-46.

Patric JH, Black PE (1968) Potential evapotranspiration and climate in Alaska by Thorntwaite's classification. USDA Forest Service, Pacific Northwest Forest and Range Experiment Station Research Paper PNW-71, Juneau, AK.

Pinay G, O'keefe T, Edwards RT, Naiman RJ (2003) Potential denitrification activity in the landscape of a Western Alaska drainage basin. Ecosystems 6(4):336-343.

PRISM Climate Group (2010). http://www.prism.oregonstate.edu/ Oregon State University, Corvallis.

Randerson JT, Chapin FS, Harden J, Neff JC, Harmon M (2002) Net ecosystem production: a comprehensive measure of net carbon accumulation by ecosystems. Ecol Appl 12:937-947.

Stieglitz M, Shaman J, McNamara J, Engel V, Shanley J, Kling GW (2003) An approach to understanding hydrologic connectivity on the hillslope and the implications for nutrient transport. Glob Biogeochem Cycles 17:1105, doi:10.1029/2003GB002041.

Strack M, Waddington JM, Bourbonniere RA, Buckton EL, Shaw K, Whittington P, Price JS (2008) Effect of water table drawdown on peatland dissolved organic carbon export and dynamics. Hydrol Proc 22:3373-3385.

Ugolini FC (1968) Soil development and alder invasion in a recently deglaciated area of Glacier Bay, Alaska. In: Trappe JM, Franklin JF, Tarrant RF, Hansen GM, (eds), Proceedings, biology of alder. USDA Forest Service, Pacific Northwest Research Station, Portland, OR, Pullman, WA, pp. 115-139.

Ugolini FC,. Mann DH (1979) Biopedological origin of peatlands in southeast Alaska. Nature 281:366-368.

USDA Forest Service (1996) Landforms of the Alaska Region classification guide. Alaska Region, December, 74 pp.

USFWS (2009) Classification of wetlands and deepwater habitats of the United States. National Wetlands Inventory website. US Dept Interior; Fish and Wildlife Service, Washington D.C. http://www.fws.gov/wetlands.

Viereck LA, Dyrness CT, Batten AR, Wenzlick KJ (1992) The Alaska vegetation classification. USDA Forest Service, Pacific Northwest Research Station, Gen Tech Rep, PNW-GTR-286.

Worrall F, Burt TP, Jaeban RY, Warburton J, Shedden R (2002) Release of dissolved organic carbon from upland peat. Hydrol Proc 16:3487-3504.

Worrall F, Burt T, Adamson J (2003) Controls on the chemistry of runoff from an upland peat catchment. Hydrol Proc 17:2063-2083.

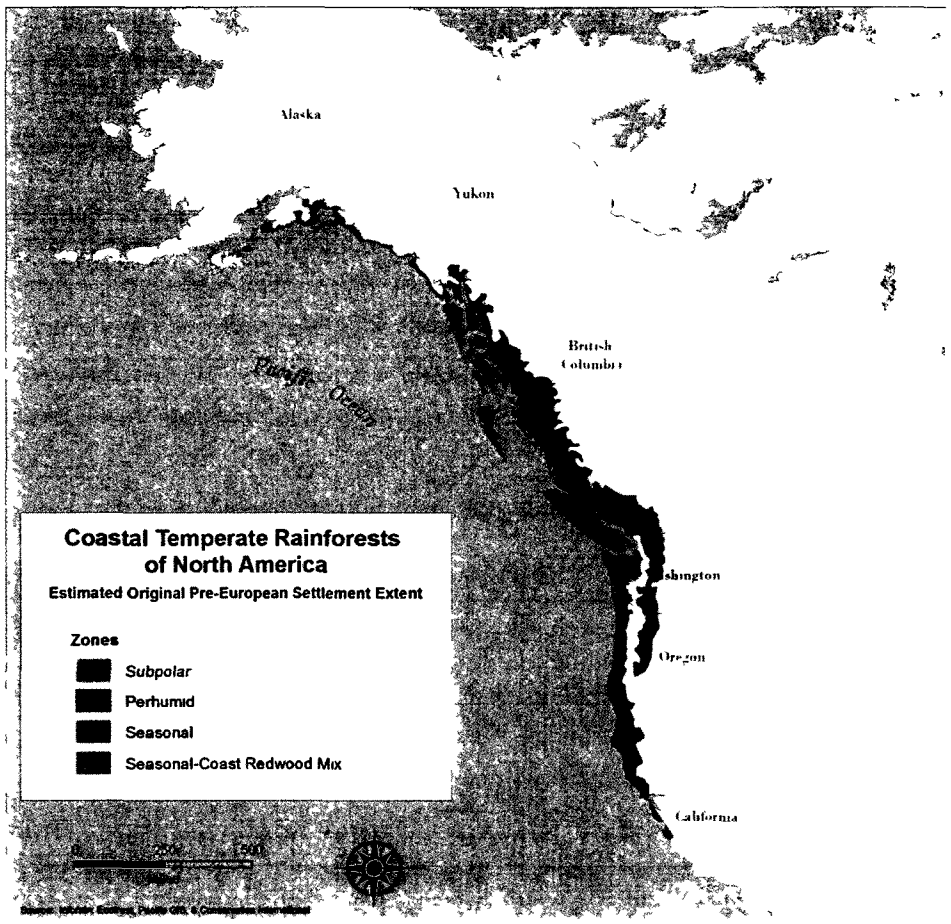

Figure 2.1. The extent of the coastal temperate rainforest along the North American Pacific coast.

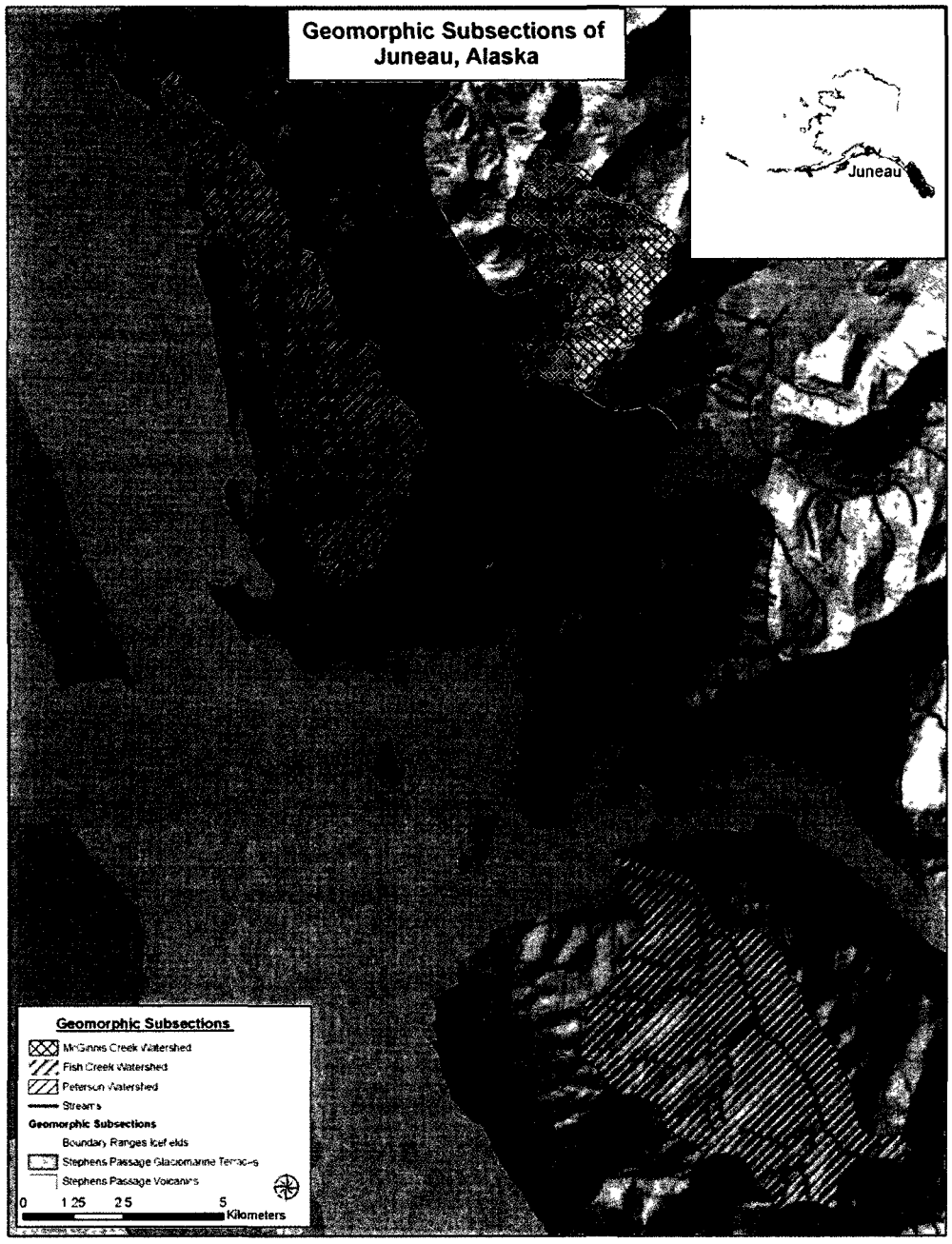

Figure 2.2. Ecological subsections and watersheds of the Juneau hydrologic observatory. Peterson Creek and McGinnis Creek watersheds include the uplifted marine terraces and Boundary Ranges Icefields, which are located on the mainland. The Eaglecrest (Fish Creek) watershed is located on the Stephens Passage volcanics located on Douglas Island.

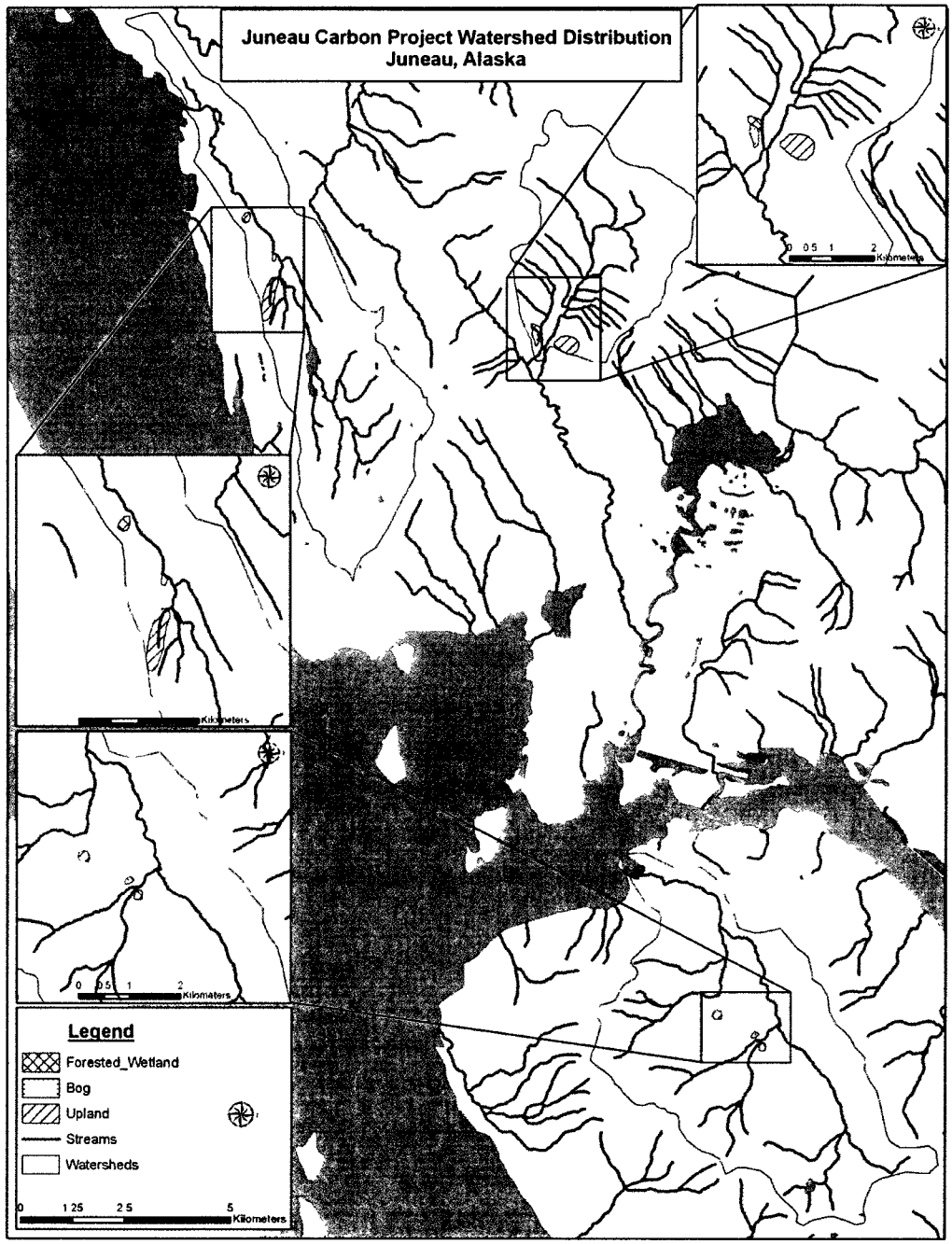

Figure 2.3. Experimental design for the Juneau hydropedologic observatory. The study includes three watersheds with catchments nested within each watershed representing each of the three hydropedologic units (sloping bog, forested wetland, and upland). The insets show locations of each hydropedologic unit (i.e. subcatchment) within the watersheds.

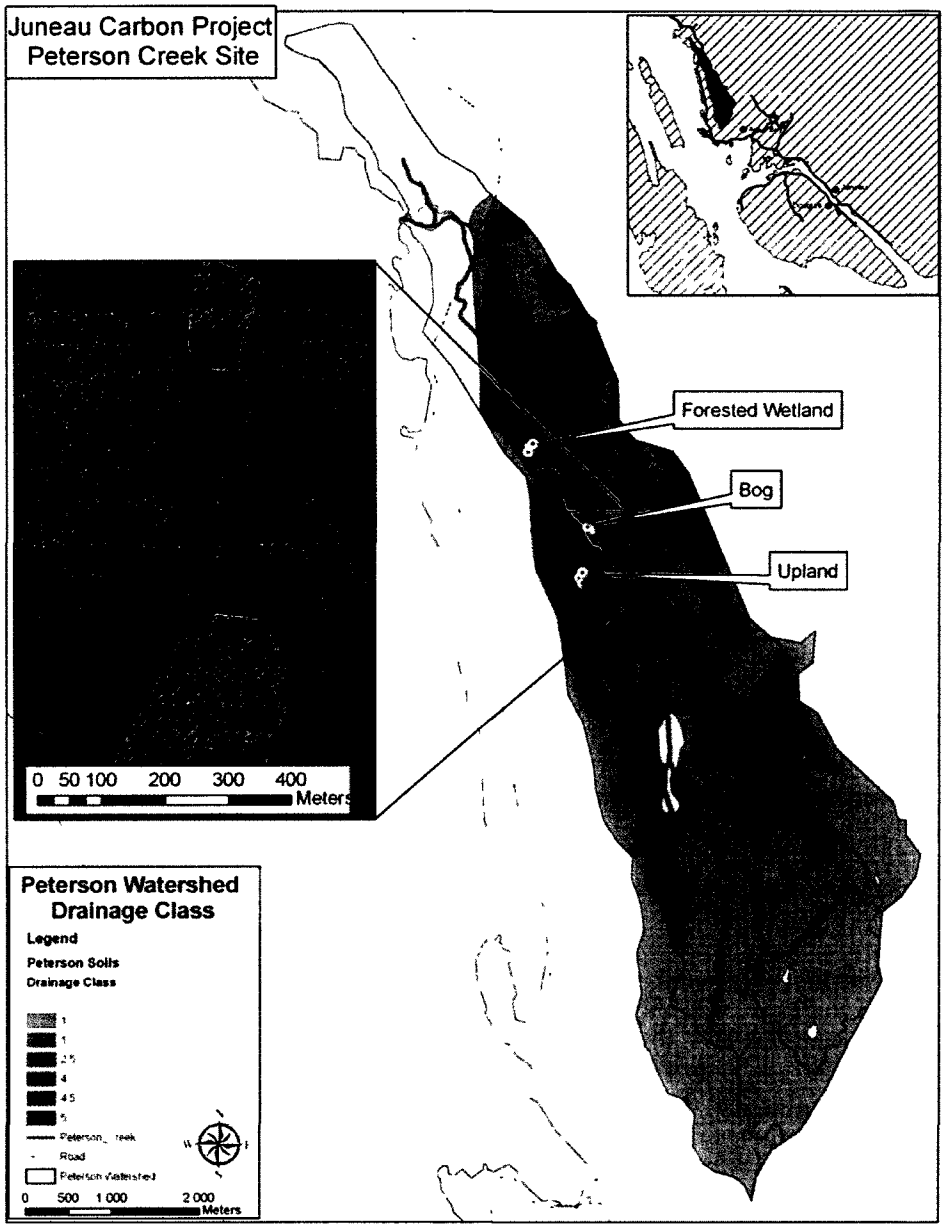

Figure 2.4. Soil drainage classes in Peterson watershed. The aerial photograph shows outlines of the upland and sloping bog hydropedologic units.

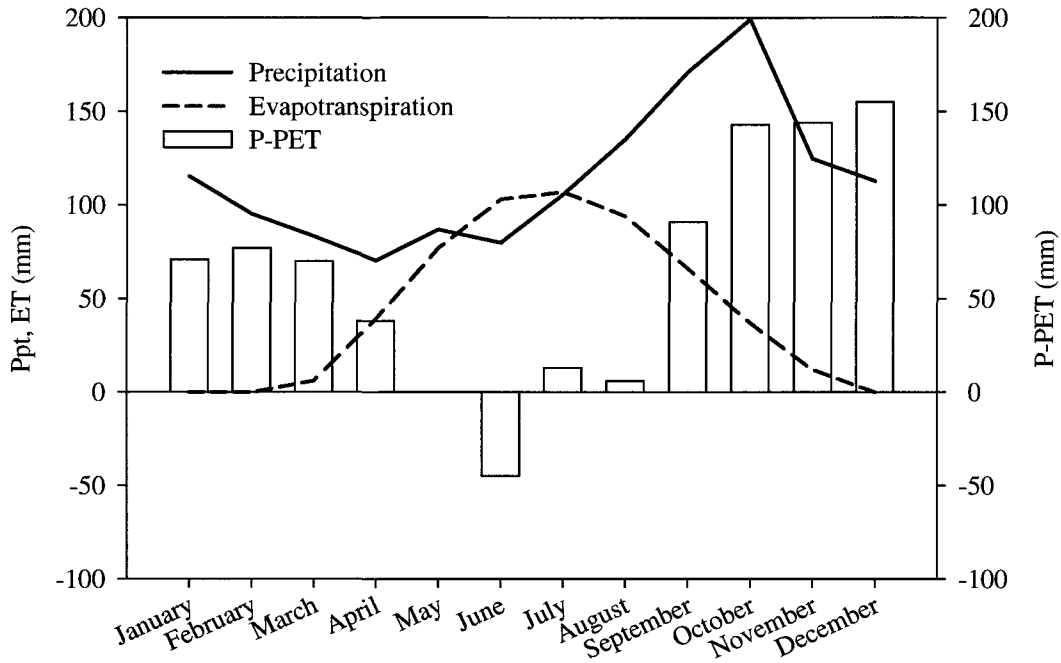

Figure 2.5. Patterns of annual precipitation and evapotranspiration in the Juneau, Alaska area. Precipitation, evapotranspiration, and the calculated moisture surplus or deficit (Precipitation [P]-Potential evapotranspiration [PET]) are shown over an annual cycle. Data were derived from the potential evapotranspiration estimates of Patric and Black (1968).

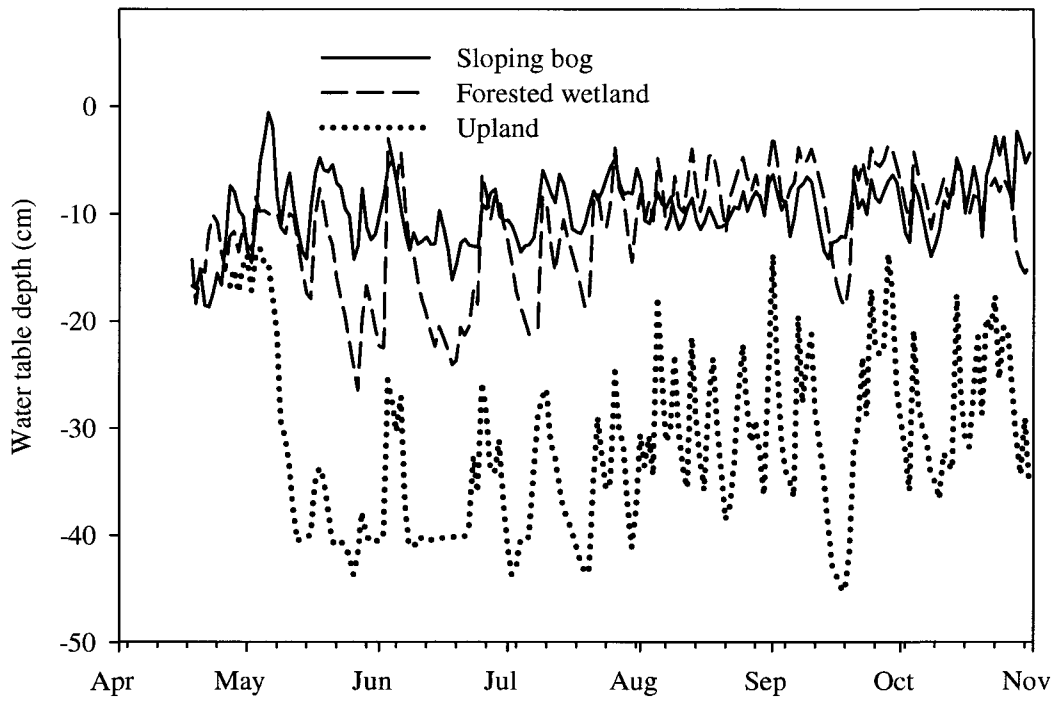

Figure 2.6. Water table fluctuations in uplands, forested wetlands, and sloping bogs. Data from three replicates of each hydropedologic units arranged in landscape classes. The period of moisture deficit can be seen most clearly in the water table depression of the forested wetland site from late May to late July.

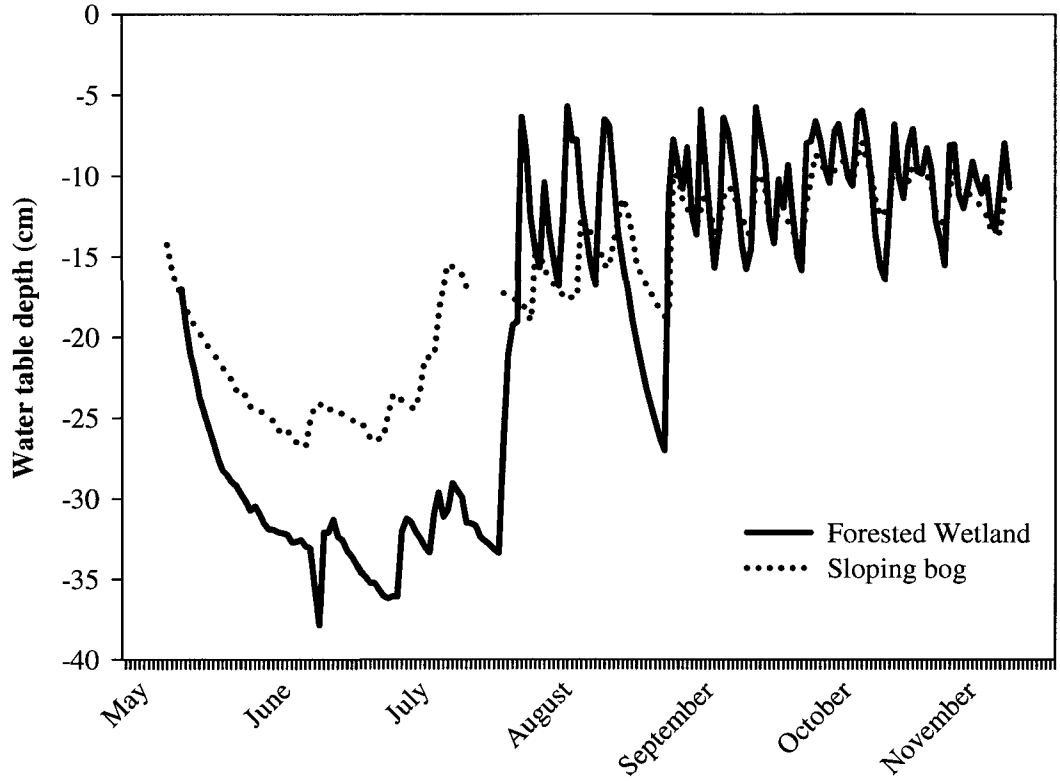

Figure 2.7. Water table depths at the McGinnis forested wetland and sloping bog. The depth and duration of aeration in the acrotelm (surface) of each soil is illustrated from May to November (from D'Amore et al., 2010).

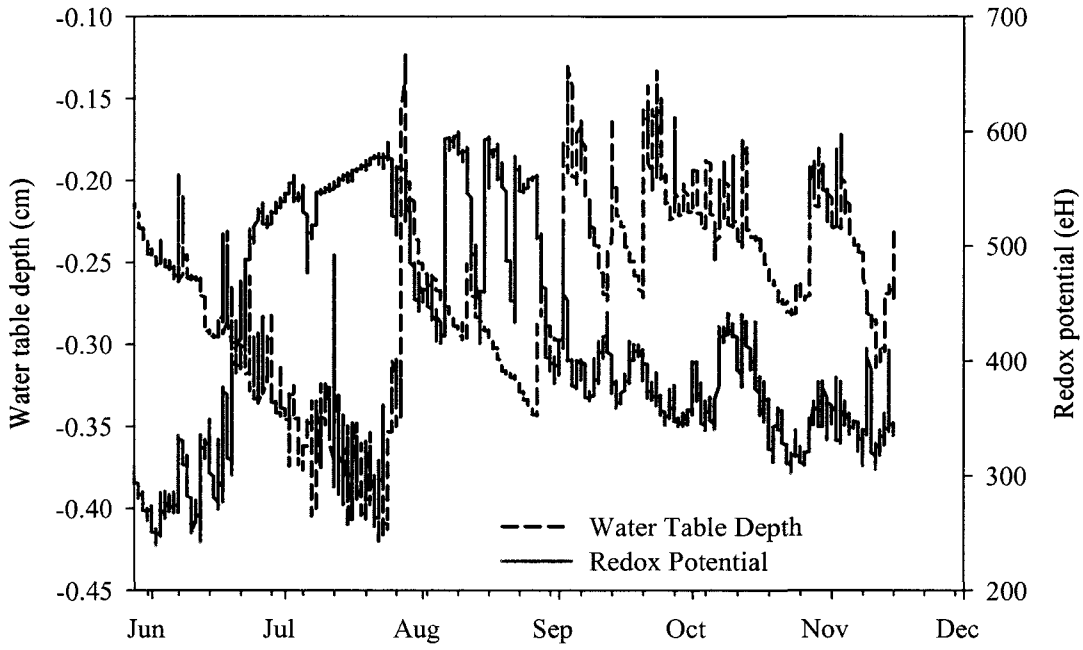

Figure 2.8. Fluctuations in water table depth and redox potential at 30 cm in the McGinnis forested wetland site during the snow-free season.

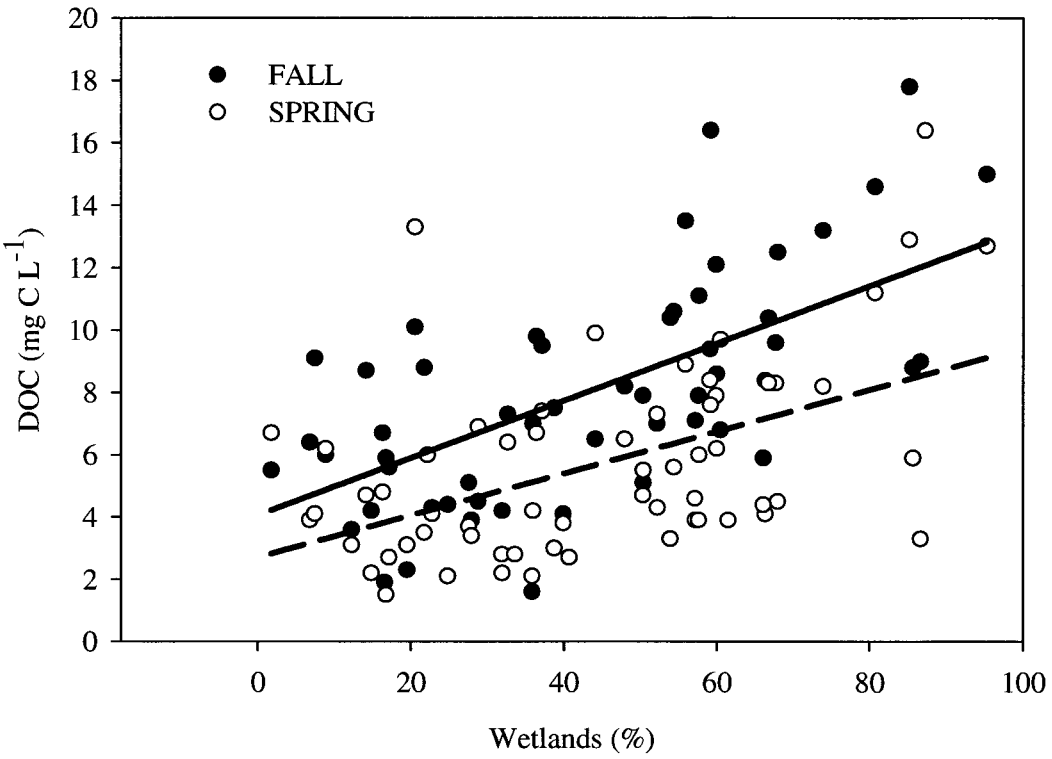

Figure 2.9. Relationship between the extent of wetlands and the concentration of dissolved organic carbon (DOC) in streams in the Alaskan portion of the northern coastal temperate rainforest. Adapted from D'Amore et al. (in review).

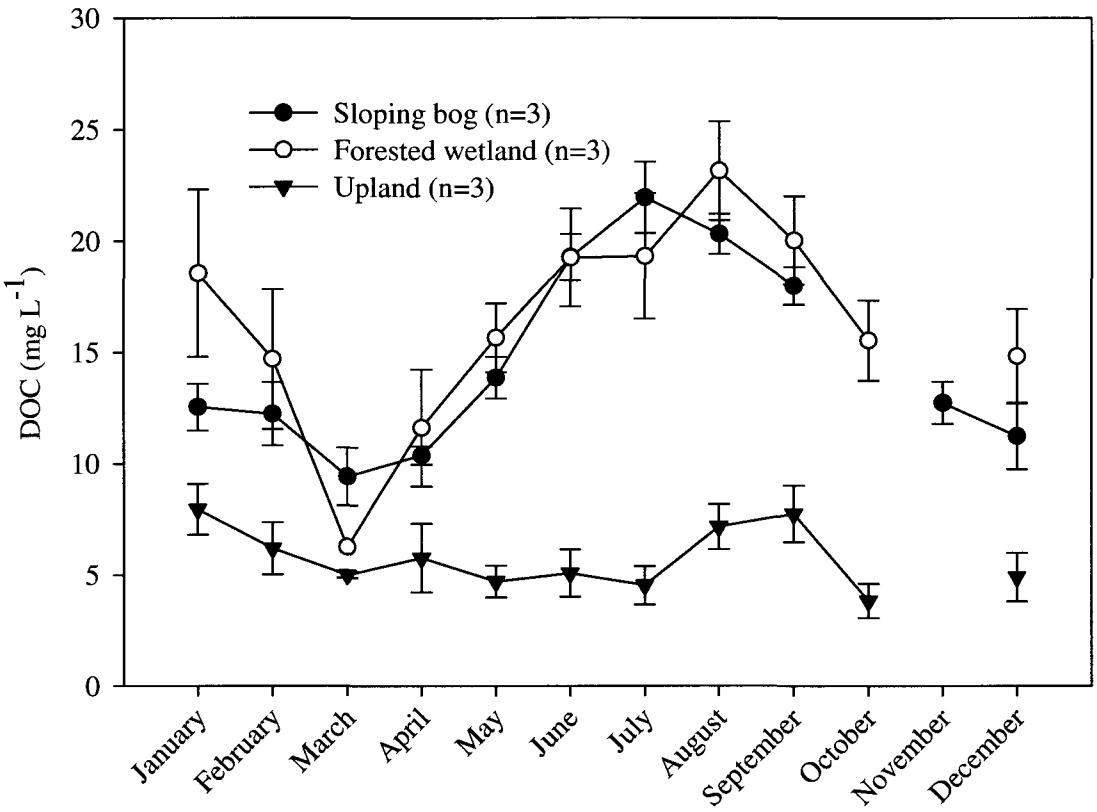

Figure 2.10. Concentrations of dissolved organic carbon (DOC) in outlet streams draining the sloping bogs, forested wetlands, and uplands in the Juneau, AK area.

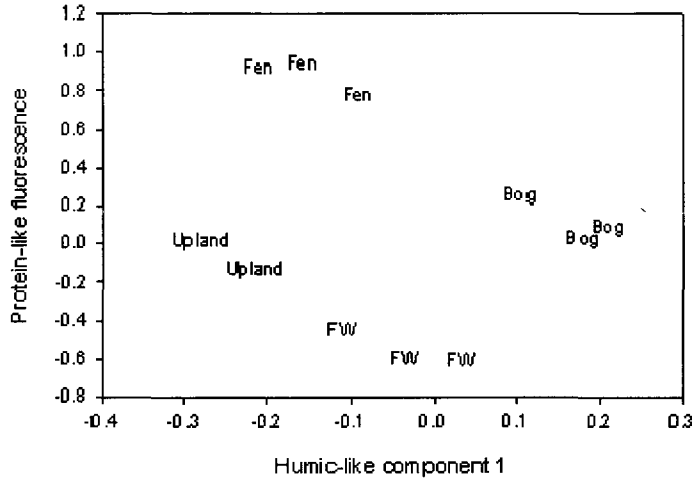

Figure 2.11. Ordination of hydropedologic units according to component composition modeled with PARAFAC analysis. See Fellman (2008; FW = forested wetland).

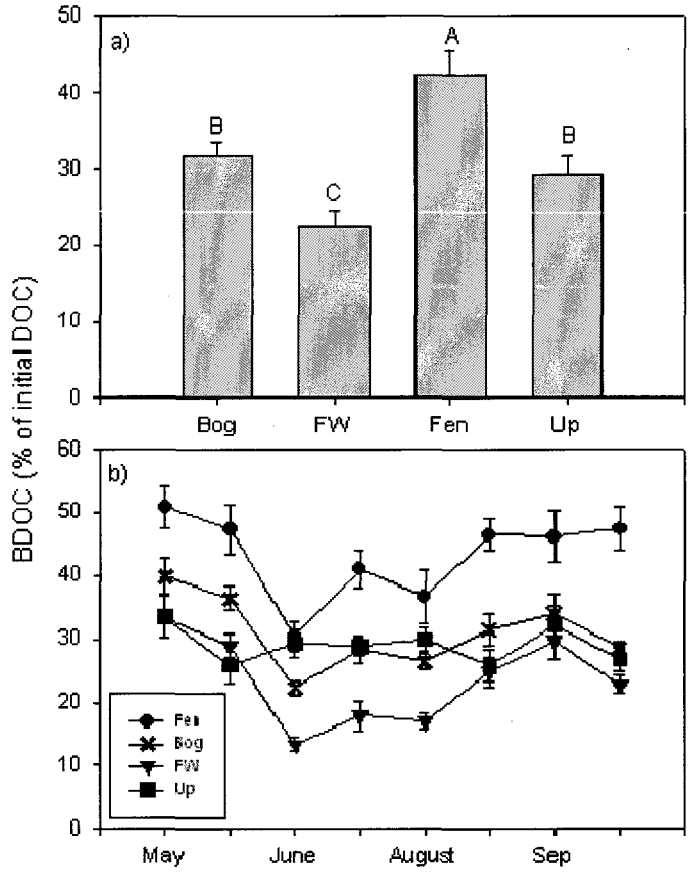

Figure 2.12. Solution biodegradable dissolved organic carbon (BDOC) in four soil types. Patterns of a) average BDOC in soil, and b) time series for the four soil types collected across the range of sample dates. Significant differences among soil types are indicated by different capital letters above the columns, error bars indicate ± 1 SE and N = 3 for all soil types. Abbreviations are: FW, forested wetland and Up, upland forest (from Fellman et al., 2008).

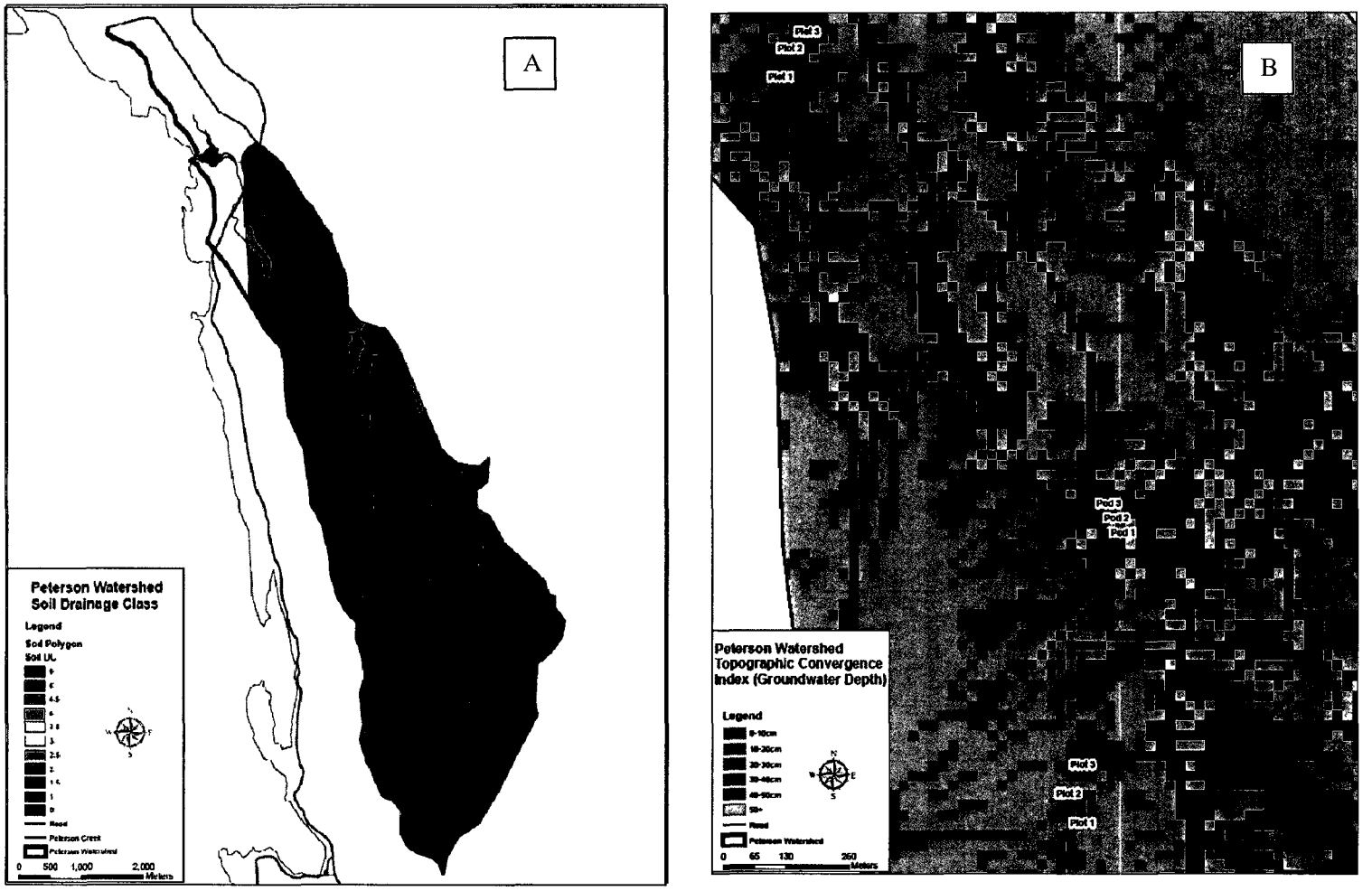

Figure 2.13. Soil drainage class and flowpaths in the Peterson watershed. Distribution of a) soil drainage classes and b) flowpaths were interpolated from a topographic wetness index.

Table 2.1. Location and attributes for sloping bogs, forested wetlands, and uplands in the Juneau, AK area.

Watershed	Hydropedologic type	Soil Type	Drainage class	NWI category*	Drainage control
Peterson	Sloping bog	Typic Cryohemist	1.0	PEM	Catotelm
	Forested wetland	Histic Cryaquept	1.5	PF	Acrotelm
	Upland	Lithic Haplocryod	5.0	UP	Bedrock
McGinnis	Sloping bog	Typic Cryohemist	1.0	PEM	Catotelm
	Forested wetland	Terric Cryohemist	2.5	PF	Acrotelm
	Upland	Typic Humicryod	4.5	UP	Bedrock
Eaglecrest	Sloping bog	Typic Cryohemist	1.0	PEM	Catotelm
	Forested wetland	Terric Cryohemist	2.5	PF	Acrotelm
	Upland	Typic Haplocryod	5.0	UP	Bedrock

*National Wetland Inventory (NWI) categories: PEM = palustrine emergent; PF = palustrine forested; UP = upland (Cowardin et al., 1979).

Chapter 3: Soil respiration rates across a hydrologic gradient in the Alaskan coastal temperate rainforest[1]

3.1 Abstract

Forest carbon budgets are of increasing concern because of their linkages with changing climate. There are very few long-term measurements of soil carbon cycles in the North American perhumid coastal temperate rainforest (NCTR), where forests and peatlands have potentially large but largely unknown fluxes of carbon, particularly through soil respiration. The easily and widely measured factors that influence the metabolism of plants and microorganisms in soils, such as temperature, moisture and substrate quality, must be coupled with a network of plot-scale measurements of soil respiration fluxes in this region in order to produce reasonable models of soil respiration flux across gradients of climate, vegetation, and soil types. We designed this study to measure soil respiration across a hydrologic gradient to quantify the influence of soil temperature and moisture on the magnitude and seasonality of carbon fluxes in the coastal temperate rainforest biome. Replicated study sites were established in three common ecosystem types (bogs, forested wetlands, and upland forest) within three coastal watersheds. Soil respiration varied significantly across the hydrologic gradient with fluxes ranging from 0.0 to 8.1 μmol CO_2 m^{-2} s^{-1} in the bog, 0.0 to 14.1 μmol CO_2 m^{-2} s^{-1} in the forested wetland, and 0.0 to 8.2 μmol CO_2 m^{-2} s^{-1} in the uplands. Soil respiration fluxes during the six-month measurement period were used to construct temperature dependent models for each ecosystem type. Temperature dependent models of soil respiration flux were fit for Lloyd and Taylor (LT) and Q_{10} models. Soil respiration in bogs was 78.9 and 84.4 g CO_2 m^{-2} y^{-1} for the LT and Q_{10} models, respectively. Forested wetlands had fluxes of 178.4 and 185.5 g CO_2 m^{-2} y^{-1} for the LT and Q_{10} models, respectively. The upland soil respiration flux was 235.3 and 250.4 g CO_2 m^{-2} y^{-1} for the LT and Q_{10} models, respectively. The soil respiration models provide a basis for estimating soil respiration fluxes across larger scales to populate global change models and reduce the uncertainty in outcomes of ecosystem response to global change in the NCTR.

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[1] D'Amore DV, Nay SM, DW Valentine (to be submitted to Biogeochemistry) Soil respiration rates across a hydrologic gradient in the Alaskan coastal temperate rainforest.

3.2 Introduction

The carbon dioxide produced by respiration and oxidation of plant roots and microbial communities associated with soil organic matter constitutes what is widely referred to as soil respiration (R_s). Soil respiration is the second largest flux in the exchange of carbon between the atmosphere and terrestrial ecosystems (Raich and Schlesinger, 1992; Schimel et al., 1994; Schlesinger and Andrews, 2000) and is the primary flux of terrestrial carbon back to the atmosphere. The general term "soil respiration" does not capture the complex interactions among plant roots, mycorrhizae, micro-organisms, and organic residues that combine with environmental influences. These biotic elements and their interactions are strongly influenced by temperature and water driving the amount of CO_2 moving from the soil to the atmosphere (Ryan and Law, 2005). Soil respiration monitoring is a key component of a comprehensive carbon cycle science program for determining carbon sequestration rates in terrestrial ecosystems. The estimation of carbon sequestration in forest ecosystems depends on constraining estimates of the major fluxes through the system. Due to the need for a better understanding of carbon sequestration, estimates of changing pools and flux rates of carbon cycling is a priority in ecosystem science. Forest carbon cycling is complex due to the diversity of biological and physical factors that interact to influence rates of accumulation or loss.

One of the densest accumulations of carbon in the world exists in the North American perhumid coastal temperate rainforest (NCTR) along the Northeast Pacific margin (Heath et al., 2011). As a result of exposure to the cold, wet conditions that limit decomposition, these coastal forests have amassed a substantial amount of biomass in both vegetation and soil carbon. The variations of topography and distribution of soil saturation on the landscape influenced the development of a diverse array of landscape types (Chapter 2). Predictions on the fate of carbon-rich forested ecosystems, such as the NCTR, are typically opposing. Such forests could act as either potential mitigation sites for the maintenance, or even enhancement, of terrestrial carbon storage. Conversely, these forests could function as a future source for increased atmospheric carbon. Concerns regarding this pool are arising as a response to predicted global warming.

Quantification of the carbon balance of this ecosystem, however, requires accurate carbon accounting models. Soil respiration measurements are needed to constrain carbon flux estimates and model carbon flux over time. The small number of soil respiration measurements in the NCTR presents a challenge to applying landscape or regional carbon flux models. The diverse array of ecosystem types also complicates the application of models constructed for coniferous

forest ecosystems. Estimates for broad regional carbon assessments are needed at present and require assessments for soil respiration rates across ecosystem types.

The challenge of developing comprehensive carbon cycle models across the NCTR is daunting. The influence of soil moisture on soil and vegetation development provides a coarse template for the entire range of ecosystem types (Chapter 2). The soil moisture gradient also corresponds to the primary influences on soil respiration: temperature and moisture. Complex terrain and high spatial variability across the forested ecosystems of southeast Alaska complicates sampling, requiring substantial spatial replication. Consequently, measurements are often constrained to distinct, accessible stand and soil conditions, Additionally, it is difficult to obtain reliable, long-term data without numerous repeated measurements, as soil respiration fluxes tend to have both high spatial and temporal variability. Given the strong association of soil respiration rates with temperature, we will test the hypothesis that the soil respiration rate will be related to differences in soil hydrologic (hydropedologic) characteristics.

We designed a project to meet the following objectives:

1. Quantify soil respiration rates across a soil hydrologic gradient and corresponding forest conditions.
2. Estimate annual flux rates for soil respiration across a highly variable spatial landscape to constrain rates for use in landscape carbon budget models.
3. Establish models of soil respiration rates for estimates of landscape soil carbon fluxes.

3.3 Materials and methods

3.3.1 Site description

We chose watersheds in three different ecological subsections as our core sites to address the higher order hydrogeomorphic control (Figure 3.1). The watersheds represent three distinct landscape types delineated by sub-section type, defined based on different lithology and dominant forms of landscape evolution (Nowacki et al., 2001). Peterson watershed, in the Stephens Passage glaciomarine terrace subsection, is primarily composed of slowly permeable glaciomarine sediments (Miller, 1973) along with bedrock outcrops that occur on moderate to low slopes. Peterson watershed represents an end-member type watershed dominated by wetlands (53% of watershed area). In contrast, McGinnis watershed is primarily composed of recently deglaciated areas within the Boundary Ranges Icefield subsection and has low wetland coverage (<5% of

watershed area). Eaglecrest watershed is composed of intrusive volcanic and sedimentary rock in the Stephens Passage volcanic subsection and represents a mix of physiographic features from alpine to lowland wetlands.

We derived soil classes that were primarily identified by hydrologic function from available surveys and inventories and delineated three sites within each of these three watershed blocks. We define these individual sites as distinct hydropedologic functional units following the concept of hydropedology proposed by Lin (2003; also see Lin et al., 2006; Chapter 2, Table 2.1). The three hydropedologic units were stratified by drainage class which was derived from the soil map unit. Histosols and inceptisols were present in both the sloping bog and forested wetland sites and the hydrologic control section was defined by the depth of the acrotelm boundary (D'Amore et al., 2010). In contrast, uplands were defined by a lack of near surface soil saturation and were located on higher gradient slopes with zones of deep soil development. The hydropedologic units are consistent with the wetland classification categories of slope bog (NWWG 1988), forested wetland, and upland (National Wetland Inventory [NWI], USFWS 2009). Wetlands are roughly mapped by the NWI, but low resolution and inaccuracies dictate that field observations of soil patterns be used to more accurately delineate the boundaries of those classes in the field. Sub-catchments that were dominated by each hydropedologic unit were delineated and a flow-gauging weir was installed to measure chemical export from each site. The distinct hydropedologic units were hypothesized to control soil hydrology and influence the overall hydrologic and biogeochemical behavior of the sub-catchment (Table 3.1).

3.3.2 Soil respiration measurement and experimental design

Soil respiration measurements were taken on collars constructed of 21 cm id PVC pipe that were permanently installed within each hydropedologic unit. The soil respiration collars were deployed in three seven-collar clusters and arranged in 2-m spacing pattern, totaling 21 collars per site. Each ecosystem site represented an experimental unit and was replicated in each watershed (n = 3 sites) for a total of 9 sites in the experimental design. Each ecosystem site type had a total of 63 collars with 21 collars per catchment type site that were measured within 1-2 days for each sample period. Soil respiration measurements were taken without drawdown using a dynamic-closed chamber procedure (Nay et al., 1994) The collars were designed to receive a portable closed-chamber field respirometer using a LiCor 820 infra-red gas analyzer (LiCor, Inc., Lincoln, NE). Soil respiration measurements were taken every 2-4 weeks during the spring,

summer, and fall over a three-year period from 2007-2009. All soil respiration measurements were accompanied by a soil temperature measurement (10-cm depth) at each collar during each measurement. Soil temperature sensors were placed at 10, 20, and 30 cm depths within a central location at each ecosystem type site, logging soil temperature hourly throughout the field season. Sensors were removed before winter, and soil temperature values of the winter months were modeled from air temperature.

3.3.3 Statistical analysis

We tested for differences in soil respiration rate among ecosystem types (sloping bog, forested wetland, upland) across time using a mixed model analysis. We accounted for the replicated measurements by using a mixed model (SAS, 2003; Proc Mixed procedure) with treatment and time as fixed effects and treatments nested within sites as a random effect.

We calculated the temperature response of CO_2 efflux rates using a non-linear least squares model (Lloyd and Taylor, 1994):

$$R = R_{ref} e^{E0\left(\frac{1}{T_{ref}-T_0}\right)-\left(\frac{1}{T_{soil}-T_0}\right)}$$

Where R is soil CO_2 efflux, R_{ref} is soil respiration (μmol CO_2 m^{-2} s^{-1}) under standard conditions (T_{ref} = 10° C), T_0 = -46.02° C (after Lloyd and Taylor, 1994), and T_{soil} is the measured soil temperature. The temperature sensitivity is the derived value for E_0 (K^{-1}). We also calculated the temperature sensitivity using the Q_{10} model as a comparison for values calculated using the Lloyd and Taylor model:

$$R = R_{10} Q_{10}^{\left(\frac{T-10}{10}\right)}$$

Where R is the soil CO_2 efflux, R_{10} is soil CO_2 efflux at 10° C, T is the measured soil temperature and Q_{10} is the temperature sensitivity.

Annual estimates for soil respiration rates were derived by applying the parameter estimates to measured or modeled soil temperatures at 10-cm depth. Soil measurements were modeled using an 11-day moving average, derived from measured soil temperatures (Zheng et al., 1993) in order to establish complete soil temperature records from 2006-2009.

3.4 Results

3.4.1 Comparison of soil respiration rates among sites

The measured soil respiration rate was influenced by both ecosystem type and season. The respiration rate was significantly related to ecosystem type ($F_{(2, 4)} = 12.85$, $P = 0.018$; Fig. 3.2), supporting our hypothesis that soil respiration rate would vary across the hydrologic gradient. The soil respiration rate was also significantly related to the time of sampling ($F_{(6, 3984)} = 46.99$, $P < 0.001$; Fig. 3.2), and the soil respiration efflux from an ecosystem type depended on the time of sampling (test of interaction: $F_{(12,3984)} = 4.11$, $P = 0.001$; Fig. 3.2). The interaction between site type and time of sampling for respiration rates among all the ecosystem types is primarily driven by the annual soil temperature trend. The magnitude and timing of peak respiration rates varied over the period of measurement each year but followed a consistent trend both within and between ecosystem types. Soil temperatures and associated soil respiration rates reach their peak values in mid-summer (Figure 3.2) and have lower rates in spring and fall. The soil respiration flux among the three ecosystem types was consistent, with the upland and forested wetland rates higher than the bogs during the growing season (Figure 3.2). The soil respiration rates in the upland sites ranged from 0.0 to 8.2 μmol CO_2 m^{-2} s^{-1} and had the highest rate in the middle of summer (mean of 2.1 μmol CO_2 m^{-2} s^{-1}; Figure 3.2). Peak mean soil respiration rates in the upland (μmol CO_2 m^{-2} s^{-1}) occurred in July 2007 and 2009. The soil respiration rates in the forested wetland were consistent during all three years of observation with peak rates occurring in July during each year (Figure 3.2). The forested wetland had a range from 0.0 to 14.1 μmol CO_2 m^{-2} s^{-1} and a mean peak of 1.3 μmol CO_2 m^{-2} s^{-1}. The rate of soil respiration efflux from the bog was consistently lower than the upland and forested wetland with consistent peak respiration rates during July and August of each year. The bog did not have as clear a seasonal signal as the upland and forested wetland and instead had a consistent range of soil respiration rates, ranging from 0.0 to 8.1 μmol CO_2 m^{-2} s^{-1} with a maximum mean value of 0.6 μmol CO_2 m^{-2} s^{-1} (Figure 3.2). The uplands and forested wetland soil respiration rates were equivalent during the shoulder seasons of spring and fall, with the uplands exhibiting the highest rates among the three sites during peak summer temperatures (Figure 3.2). The clearest separation among the sites occurred in 2007, though the general pattern was similar across the sites in each year of observation.

3.4.2 Modeling of soil respiration rates among sites

The most straightforward model of soil respiration efflux has been described through exponential functions of temperature (Lloyd and Taylor, 1994). The soil respiration rate in all three ecosystem types had a strong relationship with soil temperature (Figure 3.3), leading to favorable options for constructing temperature-dependent soil respiration models. Parameter estimates for the slope in both E_0 and Q_{10} had significant relationships with temperature and provide good model fits for predicting soil respiration efflux (Table 3.2). The average range of E_0 parameter estimates for our data (Table 3.2) were within the range of observed rates for other studies (Ruehr and Buchmann, 2010). The range of R_{ref} (R_{10}) values from the Lloyd and Taylor model and the Q_{10} model (Table 3.2) were somewhat lower than published values for temperate forests but were within a reasonable range for estimates expected in the climate of the coastal temperate rainforest where mean annual temperatures are approximately 5°C. The range of estimated Q_{10} values (Table 3.2) was similar to published values for temperate forested ecosystems (see Falk et al., 2005). The range of R_{ref} values in the Q_{10} model (Table 3.2) were also similar to published values and were very close to R_{ref} values calculated for the LT model. The coefficients in the annual models vary by year and by ecosystem type. The parameter estimates for E_0 are consistent with a decreasing activation energy from sloping bog to upland. This result is reasonable given the expected lower temperature sensitivity of vegetation and soil organic matter in saturated soils.

3.4.3 Prediction of annual respiration rates for ecosystem types

Annual soil respiration rates across the ecosystem types ranged from 72 to 247 g C m^{-2} (Table 3.3). The rate increased across the hydrologic gradient from bog to upland as a function of the gradient in predicted activation energy and reference values for soil respiration flux in the temperature-driven respiration models. The mean annual total soil respiration flux reflects the temperature sensitivity gradient of the ecosystem types and clearly separates the three types into distinct flux classes. The difference in total flux between the upland and forested wetland is 57 g C m^{-2} y^{-1}. The difference in flux between sloping bog and upland sites is approximately 156 g C m^{-2} y^{-1}. The values for Q_{10} estimates are consistently higher among all sites with 3-6% higher estimates of soil respiration flux.

3.5 Discussion

3.5.1 A regional database

Soil respiration is a key component in determining ecosystem carbon budgets. This study also fills a large void in the database for worldwide soil respiration estimates (Bond-Lamberty and Thomson, 2010), which are important for deriving estimates of carbon storage and flux both locally and at broader scales. Soil respiration ranges across a wide variety of ecosystem types and climates. Understanding the dynamics of the heterogeneous response of soil respiration can be better understood through meta-analyses. These analyses rely on comprehensive datasets with well-distributed sampling. The NCTR has some of the most carbon dense forests in the world (Heath et al., 2011), and the cycling of carbon in these coastal margin ecosystems has become the focus of attention due to the interaction with the freshwater and marine ecosystems. Aboveground carbon inventories for coastal forests in both southeast Alaska (Heath et al., 2011) and northern British Columbia provide a wide range of carbon storage in biomass, but there are no estimates for belowground carbon or fluxes to accompany the corresponding aboveground carbon pools.

3.5.2 Soil respiration variability across the hydrologic gradient

The preliminary model of soil respiration flux must be interpreted in the context of many causal factors. The soil respiration flux is influenced by temperature, moisture, plant physiology, root activity, and soil organic matter dynamics (Reichstein and Beer, 2008). One advantage of the rainforest environment is the lack of moisture limitation in the seasonal soil respiration response to temperature. In order to provide reliable estimates of soil respiration fluxes between site types through time, the experimental approach applied in our study constrained the variability of soil moisture within hydropedologic types. The generalized sampling approach we applied allows a broad application of the models for soil respiration flux across the landscape based on hydropedologic type. The use of the soil moisture status as a site factor is included in the influence on soil respiration efflux similar to explicit measurements of soil moisture. However, the predictive model is reduced to a simple temperature driven response of soil respiration flux in our approach that also captures the influence of soil saturation. The use of the hydropedologic type also incorporates the integrated influence of site factors, such as microbial community shifts, litter quality, and physiological aspects of varying plant communities.

The variation in the soil respiration rate may also be explained by two attributes of the site - overstory vegetation and soil type. The increased autotrophic potential in the uplands due to the more robust overstory canopy could increase the overall rate of soil respiration flux. The clear distinction in the soil respiration flux rates at the peak in seasonal temperature was accompanied by peaks in the upland respiration flux. Another potential mechanism is the abundance of respiration derived from the surface soils and the diffusion gradient differences in the soil types (Sanderman and Amundson, 2010). The forested wetland and bog soils are distinguished from the upland soil by a cohesive, saturated organic horizon that may impede the diffusion of gases upward in the soil profile and consequently lowering the overall flux from the soil. This phenomenon may inhibit the maximum efflux in the forested wetlands, but may also lead to sustained flux of detained gas over time. This would explain the maintenance of the respiration rates equivalent to the uplands during the shoulder seasons of ascending and descending temperatures. The sloping bog is challenged by both a lack of overstory and an increased detention of gaseous diffusion upward out of the soil surface.

The modeled values for activation energy in the LT model support our hypothesis that activation energy decreases, and associated soil respiration flux driven by temperature increases, across the hydrologic gradient. The temperature sensitivity of the sites may be due to similar vegetation, varying soil organic carbon pools, or some other mechanism. The species at both forested sites are similar with Sitka spruce (*Picea sitchensis* (Bong.) Carr), western hemlock (*Tsuga heterophylla* (Raf.) Sarg.), and mountain hemlock (*Tsuga mertensiana*) as the dominant overstory species. It seems reasonable that the trees would have the same level of sensitivity to temperature and exhibit similar physiological behavior. Another possibility for the similar temperature sensitivities is the use of labile dissolved organic carbon (DOC) compounds as microbial substrate and flux of this soil carbon intermediary. The DOC that is mobilized by the initial decomposition of soil organic matter provides a labile source for the subsequent microbial metabolism. The labile DOC would also be highly sensitive to temperature shifts in the soils. There is evidence for the loss of a large portion of the DOC measured in soil lysimeters compared to streams in a subset of soils from the sites (D'Amore et al., 2010). The measured soil respiration response may be confounded by CO_2 originating at different soil depths (Reichstein and Beer, 2008). The strong horizonation within the sampled soils creates varying pathways for diffusion which are simplified by surface chamber measurements. It is clear that diffusion plays a strong

role in the efflux of CO_2 from soils (Sanderman and Amundson, 2010) and that moisture conditions influence the diffusion of CO_2 through the soil matrix (Riveros-Iregui et al., 2008).

The wide range of values in soil respiration flux across the ecosystem types through time indicates the complex relationships among the controlling factors of soil respiration. Heterogeneity within sites is important to capture within-site variability (Rodeghiero and Cescatti, 2008), but the intensity of the within-site measurements can be uncertain (Martin and Bolstad, 2009). Our clustering of soil respiration collars addressed the objective of providing an adequate level of sub-sampling and stratification across the experimental units to provide an estimate of soil respiration flux across ecosystem types. Determining a minimum number of sub-samples and a stratification of sampling within the units is a high priority for future soil respiration monitoring in the NCTR. The use of predictive equations for generalizing parameter estimates to a finer scale of resolution for drainage classifications will provide a means for compiling landscape estimates of soil respiration that will be more tailored to actual site conditions. This approach can also be adapted to account for elevation impacts on soil respiration rates due to temperature or snowpack.

3.5.3 A framework for establishing estimates of regional soil respiration rates

Chronosequences have often been used to develop models of ecosystem development. In the NCTR, the hydrosequences, or hydrologic gradient, represents the primary variable that influences ecosystem function. In order to assess potential in initial models, the response across the hydrologic gradient has the most widespread application. The preliminary models for soil respiration rates and total carbon fluxes constrain the estimates for reasonable first approximations of plot and landscape carbon budgets. While there are many variables that provide uncertainty in the estimates of soil respiration, we designed this study to provide broad, first-order estimates of soil respiration rates across a range of ecosystem types. Reliable estimates of soil respiration rates also provide a foundation for establishing estimates of landscape soil respiration rates across the major ecosystem types in the NCTR. The variation of soil respiration rates across the gradients of ecosystem types expands the potential for capturing spatial variability in estimates that are scaled up to landscape carbon budgets. Extrapolation of the initial estimates can provide a first approximation for respiration flux across the landscape using either discrete estimates of ecosystem type or continuous associations with soil drainage class (Table 3.3). The soil hydrologic template can also be used for site and soil partitioning experiments. The

stratification of soil respiration rates by landscape type provides a framework to adjust estimates of soil respiration as more process-based knowledge is developed around ecosystem responses to temperature and moisture.

3.5.4 Modeling shifts in ecosystem carbon balance under changing environmental conditions

The ecosystem types serve as a template for inserting key information regarding systematic factors of the soil environment as explanatory factors in more complex models of predicting soil respiration efflux. The estimates in this study provide estimates of soil respiration efflux across ecosystem types at the meso-scale across the landscape. Many studies are conducted to determine the influence of soil type at the pedon, or soil-type, resolution. Despite heterogeneous soil surface conditions in this study, the experimental design allowed for the variability across a wider scale to be captured and also established a framework for designing and implementing manipulative experiments to determine changes in the trajectory of soil respiration. Soil respiration experiments are often confounded by interactions between biotic and abiotic factors (Reichstein and Beer, 2008). Characteristics of soil organic matter may play an important role in the magnitude of the response given changes in soil temperature and moisture. Stratification by dominant ecosystem captures one major source of variability driven by soil saturation patterns.

3.6 Conclusions

The soil respiration estimates and documented soil conditions in this study provide a valuable reference for larger models that incorporate a broad range of conditions across multiple ecosystem types. The overwhelming influence of soil saturation on ecosystem types across the landscape along with the close association of soil respiration rates with temperature provides a foundation for models that can be extrapolated across the landscape. Soil respiration is a key component in carbon balance models and this study provides the first predictive soil respiration models that can be applied across discrete ecosystem types for the NCTR.

3.7 Reference

Bengston P, Bengtsson G (2007) Rapid turnover of DOC in temperate forests accounts for increased CO_2 production at elevated temperatures. Ecol Lett 10:783-790.

Bond-Lamberty B, Thomson A (2010) Temperature-associated increases in the global soil respiration record. Nature 464:579-582.

D'Amore DV, Fellman JB, Edwards RT, Hood E (2010) Controls on dissolved organic matter concentrations in soils and streams from a forested wetland and sloping bog in southeast Alaska. Ecohydrology 3:249-261.

Davidson EA, Janssens IA, Luo Y (2006) On the variability of respiration in terrestrial ecosystems moving beyond Q_{10}. Glob Change Biol 12:154-164.

Davidson EA, Janssens IA (2006) Temperature sensitivity of soil carbon decomposition and feedbacks to climate change. Nature 440:165-173.

Falk M, Paw UKT, Wharton S, Schroeder M (2005) Is soil respiration a major contributor to the carbon budget within a Pacific Northwest old-growth forest? Agric Forest Meteorol 135:269-283.

Heath LS, Smith JE, Woodall CW, Azuma DL, Waddell KL (2011) Carbon stocks on forestland of the United States, with emphasis on USDA Forest Service ownership. Ecosphere 2:1-21.

Lin H (2003) Hydropedology: bridging disciplines, scales, and data. Vadose Zone J 2:1-11.

Lin H, Bouma J, Pachepsky Y, Western A, Thompson J, van Genuchten R, Vogel HJ, Lilly A (2006) Hydropedology: synergistic integration of pedology and hydrology. Water Resour Res 42:W05301, doi:10.1029/2005WR004085.

Lloyd J, Taylor JA (1994) On the temperature dependence of soil respiration. Func Ecol 8:315-323.

Martin JG, Bolstad PV (2009) Variation of soil respiration at three spatial scales: components within measurements, intra-site variation and patterns on the landscape. Soil Biol Biochem 41:530-543.

Miller RD (1973) Gastineau Channel formation: a composite glaciomarine deposit near Juneau, Alaska. Geological Survey Bulletin, Contributions to stratigraphy, USGS Report #1394, Washington DC.

National Wetlands Working Group (NWWG) (1988) Wetlands of Canada. Environment Canada, Sustainable development branch, Ottawa, Ontario, Canada. Ecological Land Classification Series 24.

Nay SM, Mattson KG, Bormann BT (1994) Biases of chamber methods for measuring soil CO_2 efflux demonstrated with a laboratory apparatus. Ecology 75:2460–2463.

Nowacki G, Krosse P, Fisher G, Brew D, Brock T, Shephard M, Pawuk W, Baichtal J, Kissinger E (2001) Ecological subsections of southeast Alaska and neighboring areas of Canada. USDA Forest Service, Alaska Region, Technical Publication R10-TP-75.

Raich JW, Schlesinger WH (1992) The global carbon dioxide flux in soil respiration and its relationship to vegetation and climate. Tellus 44:81-99.

Reichstein M, Beer C (2008) Soil respiration across scales: the importance of a model-data integration framework for data interpretation. J Plant Nutr Soil Sci 171:344-354.

Riveros-Iregui DA, McGlynn BL, Epstein HE, Welsch DL (2008) Interpretation and evaluation of combined measurement techniques for CO_2 efflux: discrete surface chambers and continuous soil CO_2 concentration probes. J Geophys Res 113, G04027, doi:10.1029/2008JG000811.

Rodeghiero M, Cescatti A (2008) Spatial variability and optimal sampling strategy of soil respiration. Forest Ecol Manage 255:106-112.

Ruehr NK, Buchmann N (2010) Soil respiration fluxes in a temperate mixed forest: seasonality and temperature sensitivities differ among microbial and root-rhizosphere respiration. Tree Physiol 30:165-176.

Ryan MG, Law BE (2005) Interpreting, measuring, and modeling soil respiration. Biogeochemistry 73:3-27.

Sanderman J, Amundson R (2010) Soil carbon dioxide production and climatic sensitivity in contrasting California ecosystems. Soil Sci Soc Am J 74:1356-1366.

SAS Institute (2003) Version 9.2, SAS Institute Inc., Cary, NC, USA.

Schlesinger WH, Andrews JA (2000) Soil respiration and the global carbon cycle. Biogeochemistry 48:7-20.

Schimel DS, Braswell BH, Holland EA, McKeown R, Ojima DS, Painter TH, Parton WJ, Townsend AR (1994) Climatic, edaphic, and biotic controls over storage and turnover of carbon in soils. Glob Biogeochem Cycles 8:279-293.

USFWS (2009) Classification of wetlands and deepwater habitats of the United States. National Wetlands Inventory website. US Depart Interior; Fish and Wildlife Service, Washington DC http://www.fws.gov/wetlands.

Zheng D, Hunt E Jr, Running SW (1993) A daily soil temperature model based on air temperature and precipitation for continental applications. Clim Res 2:183-191.

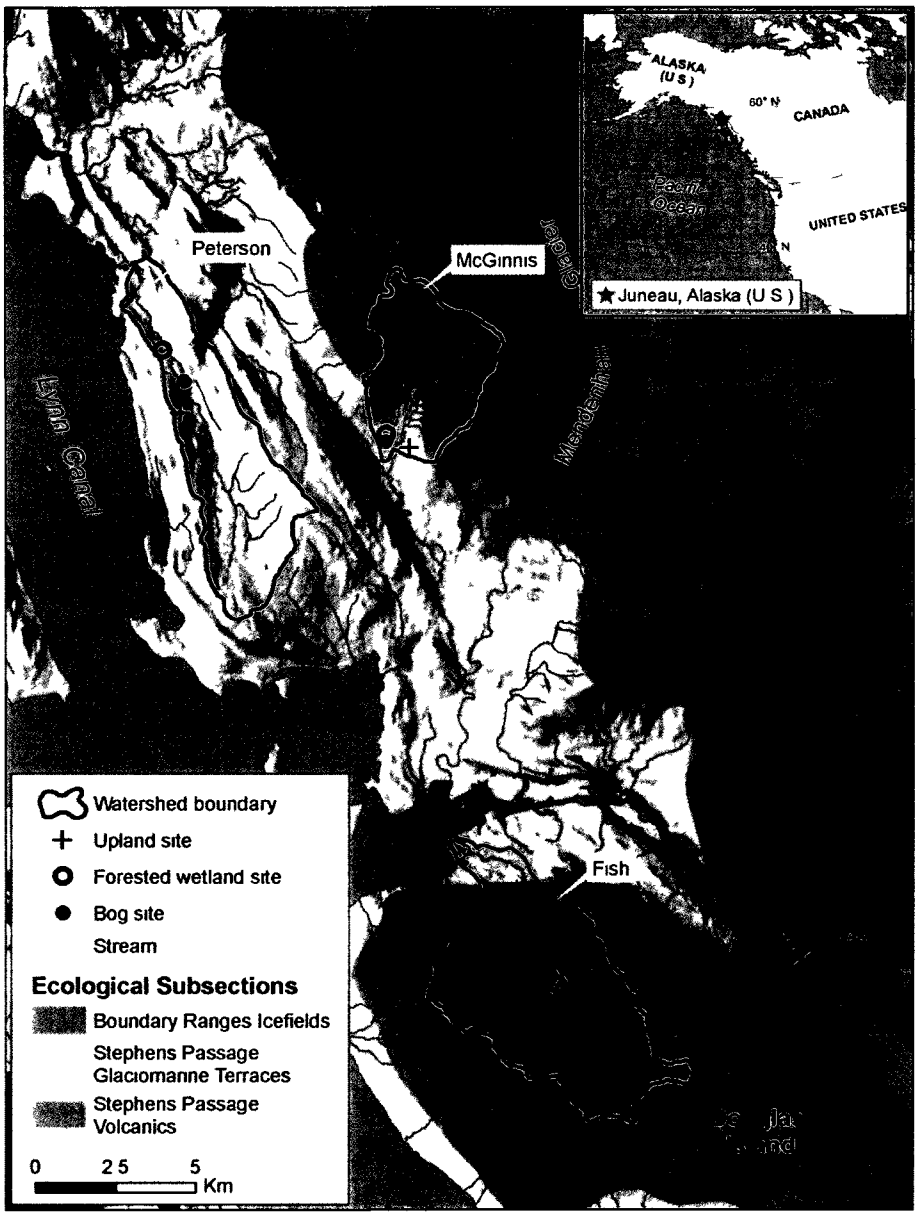

Figure 3.1. The ecological subsections and watersheds of the Juneau, AK area. The approximate locations for measurements of soil respiration rates in sloping bogs, forested wetlands, and uplands are identified within the watershed (Peterson, McGinnis, and Eaglecrest/Fish) boundaries.

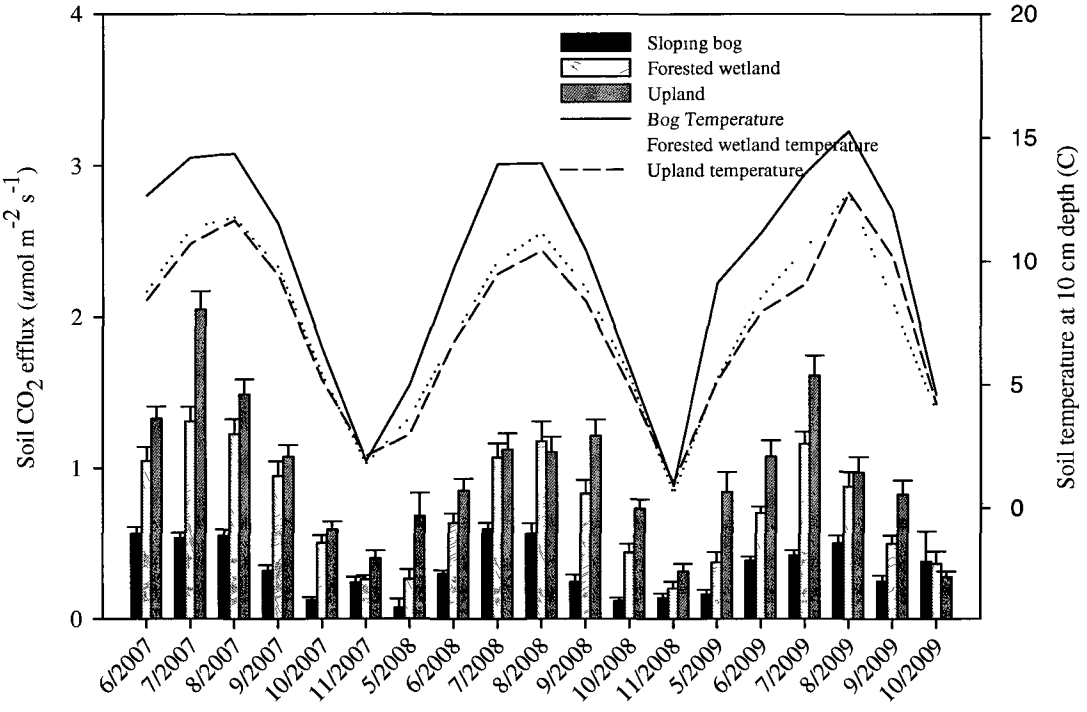

Figure 3.2. Soil respiration rates in sloping bogs, forested wetlands, and uplands in the northern coastal temperate rainforest. Measurements of soil respiration and soil temperature were taken seasonally over a three-year period. Soil CO_2 efflux rates are mean monthly rates (± SE) from soils of three replicate hydropedologic types arrayed by month and year of measurement.

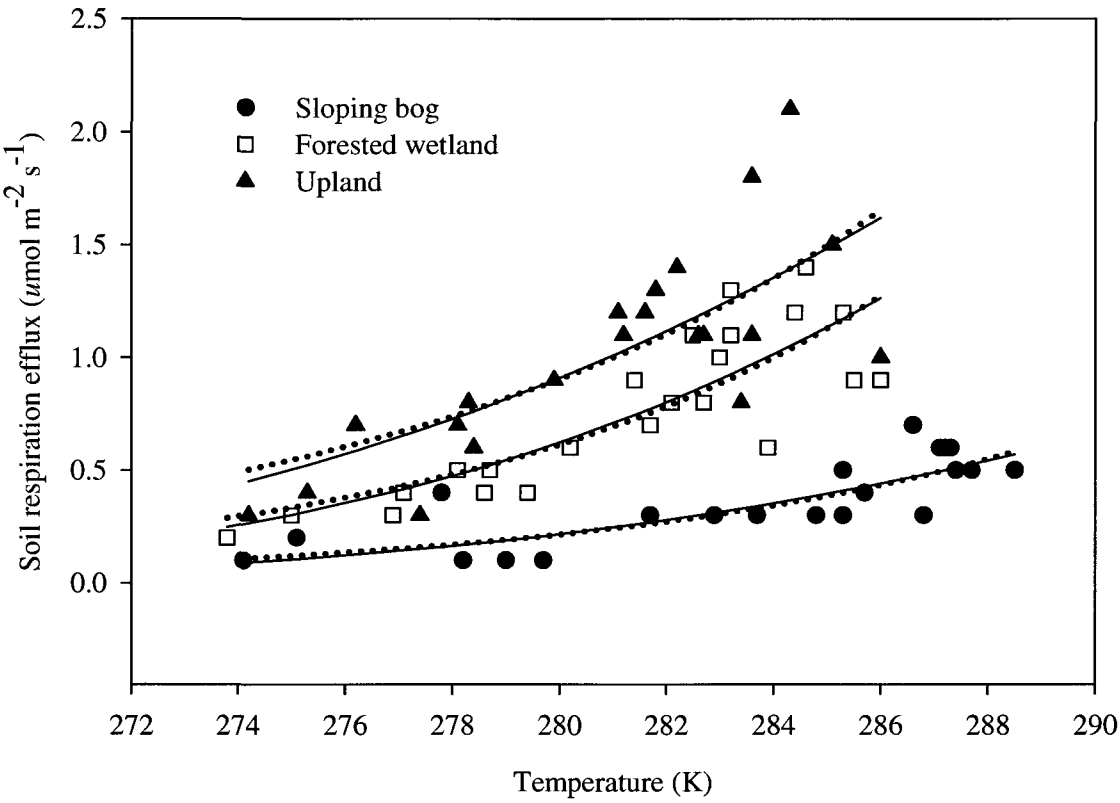

Figure 3.3. Soil respiration fluxes as a function of temperature in sloping bogs, forested wetlands, and uplands in the northern coastal temperate rainforest. Lloyd and Taylor (solid line) and Q_{10} (dashed line) model fits for each hydropedologic unit are displayed along the temperature gradient to illustrate relationships between the two models.

Table 3.1. Soil and stand characteristics in sloping bogs, forested wetlands, and uplands within three watersheds in the Juneau, AK area.

Watershed Ecosystem type	Soil subgroup	Aspect (degrees)	Elevation (m)	Slope	Average water table depth (cm)	Estimated stand age (years)	Live tree BA (m^2 ha^{-1})	Live tree volume (m^3 ha^{-1})
Peterson								
Sloping bog	Typic Cryohemist	27	112	3	-7.7	-	na	na
Forested wetland	Histic Cryaquept	55	20	5	-7.5	200	46.6	431
Upland	Lithic Haplocryod	90	163	15	-23.3	150	60.6	657
McGinnis								
Sloping bog	Typic Cryohemist	123	133	3	-6.9	-	na	na
Forested wetland	Terric Cryohemist	94	128	19	-17.2	125	44.5	387
Upland	Typic Haplocryod	280	202	20	-40.5	100	51.2	599
Eaglecrest								
Sloping bog	Typic Cryohemist	352	89	3	-7.1	-	na	na
Forested wetland	Terric Cryohemist	350	248	5	-16.5	380	38.0	248
Upland	Typic Haplocryod	343	392	15	-40.9	200	73.5	769

Table 3.2. Parameter estimates for temperature dependent soil respiration models in sloping bogs, forested wetlands, and uplands in the Juneau, AK area. Parameter estimates (± SD) for activation energy (E_0 [K^{-1}]; Q_{10}) and flux at 10° C (R_{ref}; μmol CO_2 m^{-2} s^{-1}) were fit using Lloyd and Taylor (1994) and Q_{10} models.

	Lloyd & Taylor model		Q_{10} model	
Hydropedological unit	E_0	R_{ref}	Q_{10}	R_{ref}
Sloping bog	371.79 ± 86.65	0.32 ± 0.03	3.25 ± 0.85	0.31 ± 0.03
Forested wetland	367.33 ± 62.12	0.92 ± 0.04	3.39 ± 0.71	0.90 ± 0.04
Upland	300.17 ± 74.22	1.25 ± 0.08	2.74 ± 0.70*	1.24 ± 0.08

Table 3.3. Annual estimates of soil respiration efflux in sloping bogs, forested wetlands, and uplands in the Juneau, AK area. Annual and overall total soil respiration efflux were derived from Lloyd and Taylor or Q_{10} models.

Ecosystem type	Year	Annual respiration efflux (g CO_2 $m^{-2}y^{-1}$)	
		Lloyd & Taylor	Q_{10}
Sloping bog	All years	78.9	84.4
	2006	79.8	82.6
	2007	72.0	84.9
	2008	76.2	78.7
	2009	87.4	91.3
Forested wetland	All years	178.4	185.8
	2006	176.4	183.8
	2007	176.0	182.6
	2008	169.6	176.7
	2009	191.5	200.1
Upland	All years	235.3	250.4
	2006	231.0	246.4
	2007	238.0	252.1
	2008	224.8	239.8
	2009	247.2	263.2

Chapter 4: Dissolved carbon fluxes across a hydrologic gradient in Alaskan coastal temperate rainforest soils and streams[1]

4.1 Abstract

Soil saturation and temperature are the primary factors that influence carbon cycling in soils, but the relationships between these factors vary by soil type, climate, and landscape position. Although there are extensive peatlands within North American perhumid coastal temperate rainforest (NCTR) watersheds, there is little information describing how soil saturation and temperature influence the export of dissolved organic carbon (DOC) from wetland soils in the region. To address this key information gap, we measured soil water table elevation, soil temperature, and soil and stream DOC concentrations to calculate DOC flux across a soil hydrologic gradient that included upland soils, forested wetland soils, and sloping bog soils. We found that increased soil temperature and soil water table fluctuations create conditions that promote the export of large quantities of DOC from wetland soils and relatively high amounts of DOC from mineral soils. Average area-weighted DOC flux ranged from 7.7 to 33.0 g C m^{-2} y^{-1} across a gradient of hydropedologic soil types. The total area specific export of carbon as DOC was 77, 306, and 329 kg C ha^{-1} y^{-1} from upland, forested wetland, and sloping bog catchments, respectively. The annual rate of carbon export from wetland soils in this region is among the highest ever reported. These findings highlight the importance of terrestrial-aquatic fluxes of organic carbon as a pathway for soil carbon loss within the carbon budget of the NCTR.

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[1] D'Amore DV, Edwards RT, Hood E, Herendeen, PA (to be submitted to Biogeochemistry) Dissolved carbon fluxes across a hydrologic gradient in Alaskan coastal temperate rainforest soils and streams.

4.2 Introduction

Terrestrial biological production is an important source of energy-rich reduced organic matter for stream ecosystems (Kalbitz et al., 2000; Neff et al., 2003). Soluble carbon produced in soils and exported as dissolved organic carbon (DOC) plays a role in many inorganic and organic interactions in freshwaters (Driscoll et al., 1988; Findlay et al., 1991). The transport of organic carbon to the oceans of the world is estimated at 0.4-0.9 Pg C yr^{-1} (Meybeck, 1982). A large portion of this total is from organic rich soils that export DOC organic carbon from the soil matrix to streams (Aitkenhead and McDowell, 2000).

There is a gradient of source pools for DOC in terrestrial ecosystems that is controlled by the accumulation of soil organic matter in soils. Soil organic matter accumulates primarily because decomposition is decreased by climate conditions. Decomposition is reduced by anaerobic conditions caused by water accumulation controlled by climatic gradients or topographic influence. As a result, the northern temperate forests of the world have some of the largest carbon accumulations of terrestrial carbon (Turner et al., 1995). In particular, wetlands that form due to climatic and topographic influences provide large pools of soil organic matter and serve as important sources of terrestrial carbon export to aquatic ecosystems. Decomposing plant tissues and microbial biomass turnover are the source of most soluble carbon in soils (Kalbitz et al., 2000; Michalzik et al., 2001). Low temperatures and saturated soils increase peat accumulations via reduced decomposition and accumulation of recalcitrant organic material during decomposition (Berg and McClaugherty, 2008). These biotic processes are influenced by soil temperature, soil aeration, and organic matter quality (Berg et al., 1993; Schimel et al., 1994; Bridgham et al., 1995). The retention and release of DOC in organic soils is also influenced by abiotic and biotic binding that can limit DOC solubility by the complex interaction among functional groups (Qualls, 2000; Lumsdon et al., 2005). Well-drained mineral soils are sinks for soluble carbon due to the complexation of metals such as iron and aluminum with clay particles and soluble carbon (McDowell and Wood, 1984; McDowell and Likens, 1988; Moore et al., 1992; Qualls and Haines, 1992; Kaiser and Zech, 1998). In mineral soils, the flux of soluble carbon is inversely related to this stabilization (Moore, 1989; Moore and Jackson, 1989).

The delivery of terrestrial DOC from soils through groundwater to streams varies with the seasonal balance between production and loss from solution and the potential for hydrologic transport (Boyer et al., 1997; Yano et al., 2000; Pastor et al., 2003). The solubility of DOC and

hydrologic flushing may be more critical than overall production in controlling the amount of DOC exported from organic soils (Worrall and Burt, 2008).

Quantifying DOC flux in forested catchments has relied on integrating varying source areas of production and transport (Hornberger et al., 1994). If the production of dissolved organic matter (DOM) in the soil exceeds the capacity for storage or internal consumption, DOM is available for export from the soil. One method for examining the dynamics of DOC organic carbon is to partition the landscape into functional groups according to soil type (Nelson et al., 1993; Canham et al., 2004; Creed et al., 2008). One model for partitioning the landscape is to integrate soil physical properties and chemical cycles with hydrologic export to predict the export of DOC from terrestrial to aquatic ecosystems (Eckhardt and Moore, 1990; Dillon and Molot, 1997; Gorham et al., 1998; Aitkenhead-Peterson et al., 2005). Wetlands are a major source of DOC to surface waters throughout the world (Mulholland and Kuenzler, 1979; Hope et al., 1994; Dillon and Molot, 1997; Gorham et al., 1998; Frey and Smith, 2005) and the hypothesis that carbon-rich soils or wetlands within a watershed control the concentration of DOC in surface waters is widely accepted (Gorham et al., 1998; Aitkenhead and McDowell, 2000; Mulholland, 2003; Xenopoulos et al., 2003; Frey and Smith, 2005; Huntington and Aiken, 2005).

Understanding the controls on DOC export from northern wetland soils is particularly important because nearly one-third of the estimated global soil carbon pool is stored within northern biomes (Gorham, 1991). The coastal margin of western North America from northern California, through British Columbia, to Kodiak Island in Alaska, contains the largest contiguous expanse of coastal temperate rainforest (CTR) remaining in the world (Fig 4.1). These forests have accumulated large amounts of terrestrial carbon throughout the Holocene and store more organic matter than most other north temperate forests (Gorham et al., 2007). Extensive peatlands have formed over the past 6000 years in the coastal temperate rainforest of southeast Alaska (Heusser, 1952; 1960). These peatlands store most of the terrestrial carbon in the region (Alexander et al., 1989; Leighty et al., 2006) and have the potential for substantial release and transport of DOC from terrestrial to aquatic systems (Fellman et al., 2008), but there are few studies that address fundamental controls on DOC cycling in these soils and associated streams (Emili and Price, 2006). No studies have estimated the net export of DOC from catchments within the wettest, or perhumid, region of the CTR. This region, the North American perhumid costal temperate rainforest (NCTR), includes the northern coast of British Columbia and the entire southeast Alaska panhandle (Fig. 4.1). Wetlands have been mapped across 21.7% of the

landscape of the Alaskan portion of the NCTR, compared to 5.5% in the lower 48 United States (Dahl, 2006; USFWS, 2009).

Understanding the primary control on carbon loss from the NCTR is critical to predicting the fate of the mass of soil carbon under changing climate scenarios. There is some evidence for increased DOM export from peatlands to aquatic ecosystems due to climate warming in Great Britain (Freeman et al., 2001; Worrall et al., 2004). While there is also evidence that decreased sulphate deposition may also be a factor (Evans et al., 2006), the potential that stored soil organic matter may be released as DOM through increased microbial activity and decomposition under warmer conditions (Nadelhoffer et al., 1991; Evans et al., 1999) is of concern in the NCTR. Increased DOM in the soils could be exported via shallow subsurface flowpaths, which could intensify with increased precipitation (Dalva and Moore, 1991; Qualls, 2000; Emili and Price, 2006). However, increased precipitation may also inhibit decomposition and subsequent DOM production. Therefore, understanding how DOM concentrations vary in response to seasonal changes in soil temperature and saturation is important for understanding present DOC cycling, and predicting future trends in ecosystems of the NCTR.

Carbon cycle research in forested ecosystems is a priority in managed and natural landscapes due to the important role in the carbon budget of these systems in determining sources and sinks for atmospheric carbon. Given the wet, carbon-rich landscape of the ACTR, it is reasonable to expect that the large source of DOC available for transport through streams to the ocean is derived primarily from wetlands in this region. D'Amore et al., (in review) documented the significant influence of wetlands on DOC export across the NCTR and provide predictive models for determining DOC flux from watersheds using slope. Although the total flux is much smaller than soil respiration, the movement of soluble carbon from soils is an important functional relationship between soils and streams. DOC entrained in the flow of water from wetlands provides a valuable nutrient subsidy for heterotrophic metabolism in streams of the NCTR (Fellman, 2008; 2009). Fellman et al. (2009) found that DOM extracted from wetland soils was consumed in very short distances when injected into upland streamwater.

Quantifying the total export of DOC and identifying the controls on soluble carbon production and export from soils is necessary to understand combined terrestrial and ecological functions. We compared measurements of soil water table level and soil temperature across replicated soil end-member units to explain patterns of DOC flux in watersheds that spanned the hydrologic gradient in the NCTR. We combined measurements of streamwater DOC

concentration and estimated annual discharge to quantify the flux of DOC from discrete sub-catchments of the NCTR.

4.3 Materials and methods

4.3.1 Site description

Research was conducted in three watersheds near Juneau, AK that are part of a long-term carbon flux study in the North American Carbon Program (NACP; http://www.nacarbon.org/nacp/; Figure 4.1). We chose watersheds in three different ecological subsections as our core sites to address the higher order hydrogeomorphic control on ecosystem functions (Figure 4.1). The watersheds represent three distinct landscapes delineated by sub-section type dominated by different lithology and dominant forms of landscape evolution. (Nowacki et al., 2001). Peterson watershed, in the Stephens Passage glaciomarine terrace subsection, is primarily composed of slowly permeable glaciomarine sediments (Miller, 1973) along with bedrock outcrops that occur on moderate to low slopes. Peterson watershed is also a wetland-dominated watershed (53% of watershed area). In contrast, McGinnis watershed is primarily composed of recently deglaciated areas within the Boundary Ranges Icefield subsection and has low wetland coverage (<5% of watershed area). Eaglecrest watershed is composed of intrusive volcanic and sedimentary rock in the Stephens Passage volcanic subsection and is a mix of physiographic features that include alpine, productive forest and wetlands.

We derived soil classes that were primarily identified by hydrologic function from available surveys and inventories and delineated three sites within each of these three watershed blocks. We define these individual sites as distinct hydropedologic functional units following the concept of hydropedology proposed by Lin (2003; also see Lin et al., 2006; Chapter 2, Table 2.1). The three hydropedologic units were stratified by drainage class which was derived from the soil map unit. Histosols and inceptisols were present in both the sloping bog and forested wetland sites and the hydrologic control section was defined by the depth of the acrotelm boundary (D'Amore et al., 2010). In contrast, uplands were defined by a lack of near surface soil saturation and were located on higher gradient slopes with zones of deep soil development. The hydropedologic units are consistent with the wetland classification categories of sloping bog, forested wetland, and upland. Wetlands are roughly mapped by the National Wetland Inventory (NWI), but low resolution and inaccuracies dictate that field observations of soils patterns be used to more accurately delineate the boundaries of those classes in the field. Sub-catchments that

were dominated by each hydropedologic unit were delineated and a flow-gauging weir was installed to measure chemical export from each site. The distinct hydropedologic units were hypothesized to control soil hydrology and influence the overall hydrologic and biogeochemical behavior of the sub-catchment (Table 4.1).

4.3.2 Soil description and sampling

Soil profiles were described a sloping bog, forested wetland, and an upland site within each of the three study watersheds from a representative pedon. The pedons were then classified to provide specific soil attributes for each site (Soil Survey Division Staff, 1999; Table 4.1). The soil was characterized to either an impermeable layer (till or bedrock) or to a depth of 2 m (Table 4.2). A modified syringe adjusted to 5-ml capacity and a 100-mesh sieve was used to calculate rubbed and unrubbed fiber content (Lynn et al., 1974; D'Amore and Lynn, 2002). Samples from each soil horizon were dried and ground to pass through a 177 μm (80 mesh) sieve for total carbon and nitrogen analysis by combustion on a CHN analyzer. Soil pH was determined on moist 2.5 cm^3 subsamples in 4 ml of 0.01 m $CaCl_2$. Bulk density was determined by measuring pore volume on 125 cm^3 samples carved from soil pit faces with sharp knives for organic material or with a volumetric core for mineral material. Organic matter particle density was assumed to be 1 g cm^3 for determination of bulk density. Gravimetric water content was determined on three replicate 50 g subsamples of soil dried at 105°C to a constant weight.

Water tables were measured using pressure transducers placed in two 3.2 cm diameter PVC tubes installed to 50 cm depth near the soil sample locations. Water table levels were recorded every four hours and averaged to daily or monthly values at each site. Soil temperature was recorded with thermistors at 10, 20, and 30 cm depths. Soil temperatures for missing data periods were supplemented with modeled soil temperature measurements from air temperature following the method of Zheng et al. (1993).

4.3.3 Soil and stream water collection and analysis

Soil pore water samples were taken from two piezometers installed at 25 and 50 cm depths. Soil pore water was composited from the two piezometers at each depth during each collection every 1 to 3 weeks, yielding one sample from each depth for each site. Streamwater samples were taken from the tributaries draining the sub-sites every 1 to 3 weeks during 2006-2007. Soil solution and streamwater samples were field-filtered using pre-combusted, Gelman A/E glass fiber filters and stored in the refrigerator until analysis within one week of collection.

DOC concentrations were analyzed using high temperature combustion, Shimadzu TOC/TN-V analyzer.

4.3.4 Stream discharge

Flow gauging weirs were installed in outlet streams draining each sub-catchment site. The weirs were positioned to capture most of the water that flowed from the unit. Each weir was equipped with a pressure transducer that measured water height in the flume. Transducers were corrected for atmospheric pressure changes by barometric pressure transducers at the site. In the case of missing or erroneous barometric data, atmospheric readings from a NOAA barometric station was used for corrections. Continuous flow measurements were made at the flumes unless the flume was frozen.

4.3.5 Calculation of streamwater DOC flux

Streamwater DOC flux from each end-member subcatchment was calculated from continuous discharge measurements combined with intermittent streamwater samples of DOC concentration. Concentration-discharge relationships and estimation of DOC flux was calculated using the load estimator program LoadEst (Runkel et al., 2004). Data input and output was facilitated by use of the Loadrunner program (Raymond and Saiers, 2010; http://environment.yale.edu/loadrunner/loadrunner/readers/discussion.html). The flux of DOC was calculated with an adjusted maximum likelihood estimation through a non-linear regression which was derived from the measured DOC concentration and measured stream discharge as dependent and independent variables (see Runkel et al., 2004 for details). Estimates of daily flux were calculated from measured discharge and the concentration discharge relationship computed through the maximum likelihood estimation. Daily estimates were summed for each watershed, and area-weighted fluxes were derived by distributing the total annual flux over the watershed area of the catchment.

4.3.6 Statistical analyses

We tested the influence of hydropedologic sub-catchments (sloping bog, forested wetland, upland) on average monthly soil water table depth. We accounted for the measurements replicated in time by using a mixed model analysis (SAS, 2003, Proc Mixed procedure) with treatment and time as fixed effects and treatments nested within sites as a random effect. There were no departures from the assumptions in the analysis (i.e. normality, error variance) so we

analyzed untransformed data. We chose a compound symmetry (CS) covariance structure after evaluating various other covariance structure options. Soil DOC concentrations as a function of temperature were described by fitting the Lloyd and Taylor (1994) (i.e. Arrhenius) functions to the measured DOC concentrations across time in the ecosystem units. The relationship between water table depth and the concentration of DOC in streams was evaluated with a one-way ANOVA model with ecosystem type as the class variable using a Tukey-Kramer post-hoc means separation test. A multiple regression model was applied to the combination of water table depth and soil temperature to test the relationship with streamwater DOC concentrations and water table depth and temperature variables. All ANOVA and regression analyses were done using Sigmaplot's statistical package (Sigmaplot, V.11).

4.4 Results and discussion

4.4.1 Soil pedologic development and features related to soil organic matter storage

The low temperatures and reducing conditions due to prolonged saturation in the soils created by the regional climate promote soil carbon storage across all soil types. The soils are a product of different suites of soil formation and are in the Inceptisol, Spodosol, and Histosol soil orders. They include moderately decomposed Hemists, a well decomposed Saprist, a poorly-drained Aquept, and moderately well-drained Cryods (Table 4.1). Soil carbon stocks are substantial in all soil types that increase from uplands to forested wetlands and bogs (Table 4.2). The depth and density of the soil organic matter are highest in the wetland soils where carbon storage ranges between 159 and 614 kg C m^{-2}. The upland soils have a mean carbon content of 157 kg C m^{-2}, with a range from 145 to 174 kg C m^{-2}. The main storage pool for carbon in all the soils is in the surface organic horizons that have low bulk densities but a high carbon percentage (Table 4.2). There are varying climate patterns across the NCTR, but the overwhelming influence of low temperatures and high rainfall lead to the development of wetland soils with large stocks of carbon. The carbon storage values measured in the wetland soils of this study are similar to the range of carbon storage found in similar soils throughout the NCTR (Alexander et al., 1989; D'Amore and Lynn, 2002) as well as similar northern climates with peatlands such as the United Kingdom (Harrison et al., 1995).

In the cool, wet climate of the NCTR, topographic position exerts an overwhelming influence on patterns of organic matter accumulation due to soil saturation (Neiland, 1971). The soil types are also influenced by the type of parent material and feedback from vegetation that

leads to distinct horizonation. The organic horizons are distinguished by patterns of decomposition and drainage. Soil development among the sampled horizons reflects the influence of lateral topographic gradients and the associated vegetation response. The vertical soil horizonation of the peatlands (i.e. forested wetlands and sloping bogs) is primarily the result of the accumulation of plant detritus above permanently saturated humified material. The key difference is the depth of the humified material and the depth to the water table. The well-drained upland soils also exhibit a vertical horizonation, but the organic horizon is unsaturated and meets the soil taxonomic criteria for a folistic epipedon in some cases (USDA Forest Service, 1997). These upper organic horizons contain relatively dense carbon stores (Table 4.2). The moisture excess (P>PET) promotes the saturation of soils on progressively lower slopes as water accumulates and soil water routing becomes limited, thus resulting in a reducing condition that favors organic matter accumulation.

4.4.2 Soil hydrodynamics

The expression of the soil horizonation leads to distinct soil hydrodynamics among the measured pedons. Soil saturation exerts a strong influence on forest stand vigor in the NCTR, and soil hydropedologic units have previously been distinguished according to their hydric soil status to predict timber volume classes (Caouette and DeGayner, 2005). However, hydric soil status is a binary system and does not provide a range of soil water table depths. In this study, the depth to water table varied significantly among the hydropedologic units ($F_{(2,4)} = 23.80$, $P = 0.006$; Figure 4.2) and varied by season ($F_{(12,68)} = 4.71$, $P < 0.001$; Figure 4.2). The significant difference in the water table depth among the hydropedologic units provides a greater scope for interpretation of water table fluctuations and associated response of vegetation and carbon. The hydrologically conductive surface horizon in the sloping bog and forested wetland extended to 15 and 25 cm depths, respectively. These organic horizons consist of freshly deposited material along with moderately humified organic material (Table 4.2) that permitted the drainage of the soil water during the early part of the growing season, but became saturated as precipitation increased in the fall (Figure 4.2).

However, soil saturation, as defined by hydric soil status, does not readily apply to specific biogeochemical processes or functional responses within a specific soil type. For instance, though soil saturation leads to organic matter accumulation and the development of forested wetlands and sloping bogs, the timing and duration of saturation is not captured by this

distinction. The forested wetland and sloping bog sites have limited drainage owing to their gentle slopes, but still exhibited a moisture deficit in the early to mid-summer associated with increased water demand by plants coinciding with reduced precipitation. The water table drawdown was a consistent feature in the soils throughout the period of observation in the present study and in related studies of soil moisture patterns (D'Amore et al., in prep.; D'Amore et al., 2010). The drawdown is related to the amount of conductive organic material located at the surface of the soil known as the acrotelm, and the more slowly conductive peat lower in the profile known as the catotelm. This two tiered peat system is associated with hydrologic patterns in the sloping bog and forested wetland at McGinnis Creek (D'Amore et al., 2010). The consistent and statistically significant pattern of soil saturation among the replicated hydropedologic units confirms that the patterns of soil saturation are a persistent feature in the soils.

Streamwater discharge is closely correlated to water table depth within the soils (Fig. 4.3). Stream discharge increases when water tables rise, with little or no time lag, indicating that there is very little detention storage in the soils and precipitation is quickly translated into streamflow. The lack of saturation excess overland flow is characteristic of soils in the coastal temperate rainforest due to the permeability of surface organic horizons and the slope of the landscape (Bishop, 1955; 1968). Wetlands are normally thought to retain water and introduce a lag between rain and discharge. Organic soils, such as the sloping bogs and forested wetlands, have a highly conductive acrotelm overlying a slowly permeable catotelm, which results in soil water flow in the upper part of the soil (Holden and Burt, 2003) and longer term storage in the less permeable catotelm. The bogs and deep organic horizons of the forested wetlands certainly serve as long term water reservoirs and lose water very slowly. The surface soils are much more responsive to rainfall events and readily convey water rapidly to nearby water courses (Bishop, 1968). The upland soils have a high soluble carbon sink capacity in the spodic horizon due to the downward drainage gradient in the soil horizon and the DOC being stabilized by forming complexes with Al and Fe. These soils often overlie bedrock or till, which serves as an aquitard for water movement out of the soil.

The strength of correlation between soil water table and stream discharge decreased from uplands, to forested wetlands and sloping bogs. The association of soil water fluctuations appears to indicate the ability of the soils to detain and store water consistent with the slope (Table 4.1) and the soil moisture storage in the peatlands. Upland soils rapidly drain water from the profile

and translate the excess water into streamflow, while forested wetlands and sloping bogs detain and store more precipitation. The water table fluctuations are significantly related to the streamwater discharge among all the soil types highlighting the large amount of precipitation that is conveyed to stream drainages after storms (see Fellman et al., 2009).

4.4.3 Production and fate of soluble organic carbon in soils

The stored soil carbon pool derived from plant material can be respired as CO_2 or converted to soluble organic compounds, mostly as organic acids, which can be transported or respired to CO_2. In surface soil layers, this carbon pool is actively decomposed due to warmer, aerobic conditions compared to the subsurface horizons (D'Amore et al., 2010). Soil temperatures increase in spring and peak in late July and August at all sites (Fig. 4.4). The temperatures have distinct shoulder seasons of ascending (spring) and descending (summer) temperatures around the peak. Soil solution concentrations in the surface horizons peak in July and August during the peak temperature period. Subsurface DOC concentrations peak in May and decrease in concentration throughout the summer and fall. There is also a clear distinction in solution DOC concentration in the subsurface between the uplands and wetlands in which low DOC concentrations appeared at 25 and 50 cm respectively in April.

Soluble carbon derived from the soil carbon stock is strongly related to the seasonal temperature and soil saturation cycles (Fig. 4.5). Soil solution at 25 cm depth across all sites was significantly related to soil temperature ($F_{(1,83)} = 44.30$, $P < 0.001$) and soil water table depth ($F_{(1,85)} = 7.14$, $P = 0.009$). A combination of both water table and soil temperature yielded a model for soil solution concentrations at 25 cm depth ($F_{(2.84)} = 24.89$, $P < 0.001$) based on the average monthly values for the inputs across all sites. Soil solution DOC concentrations and temperature were used to construct temperature dependent exponential functions (Lloyd and Taylor, 1994). The significant model fits (Table 4.3) indicate that the temperature dependent model provides a good approximation of DOC production in the soils. Parameter estimates in this model provide a means to predict soil solution DOC concentrations as indicated by a baseline value for DOC in soil solution at a reference temperature (10° C) and the response of DOC production related to activation energy of the soil biological communities interacting with the soil organic matter store.

4.4.4 Streamwater dissolved organic carbon fluxes

The delivery of DOC to streams from the sub-catchments followed the pattern in the soil matrix where temperature driven DOC produced in the soils moved into streams when soil water

tables were high. Streamwater DOC concentrations were significantly different among soil types (Fig. 4.6) with concentrations in streamwater draining the wetland soil types 2-3 times the concentrations of water draining the uplands. The combination of seasonal soil DOC production and increased discharge across all the ecosystem types culminates in a strong fall flux of DOC from the wetland catchments and a slight increase in the upland catchments (Fig. 4.6). Temperature and water table depth were significant predictors of stream DOM (multiple regression: $F_{(2,17)} = 44.84$, $P < 0.001$). DOC flux patterns are also related to the discharge relationships in the catchments as peak fluxes are reached later than peak DOC concentrations, but associated with peak discharge in the fall.

The export of DOC from catchments has been described as resulting from the effect of precipitation and temperature on the dynamics of DOC export (Clair et al., 1996). However, temperature alone may not be the best predictor of DOC export because warm temperatures that produce peak soil solution DOC concentrations may not coincide with effective hydrologic transport (Worrall et al., 2004). Much of the DOM transported from northern temperate terrestrial ecosystems to streams is derived from fresh organic material produced by plant detritus (Schiff et al., 1997). Because the quality and amount of recent detritus varies widely among peatbogs, forested wetlands, and upland forest, the ecosystem type combined with temperature and soil saturation patterns will likely influence the ultimate flux of DOC from a catchment (Bengtson and Bengtsson, 2007). DOC entrained in the flow from the NCTR catchments appears to be more closely related to the balance between production, consumption, and transport of soluble carbon in the soils. Concentrations of soluble carbon in the soils increase during the summer and peak in conjunction with peak soil temperatures. Intermittent storms flush stored DOC out of surface horizons to streams and produce a characteristic fluctuating curve of streamwater DOC concentrations associated with stormwater over the short-term (Fig. 4.6). The seasonal pattern of DOC concentration in streamwater is consistent with the peak coincident with fall increases in precipitation and the filling of the entire soil horizons at the sloping bogs and forested wetlands, while the uplands remain consistent in the flux of DOC from the ecosystem types (Fig. 4.6).

4.4.5 Watershed model for dissolved organic carbon flux

Annual DOC flux budgets were calculated from the replicated ecosystem types. The total annual export from the sub-catchments revealed the strong seasonal pattern of DOC export as well as the influence of hydropedologic unit (Fig. 4.7). The Total annual export of DOC from all

watersheds exceeded 100 kg ha^{-1} (Table 4.4). This value exceeds most reported fluxes for comparable systems (Clair et al., 1994; Hope et al., 1997; Canham et al., 2004). The range of export upward to over 400 kg ha^{-1} (Table 4.4) reveals the potential of the landscape to re-distribute carbon from the terrestrial ecosystem in quantities far from those normally experienced in most ecosystems. A comparison of average annual flux from the ecosystem types indicates that the estimates are closely associated with values from the literature for low runoff:DOC in the uplands and high runoff:DOC in the sloping bogs and forested wetlands (Fig. 4.8). The annual output of DOC from upland watersheds supports the concept that the lateral fluxes of water promote export of DOC. The average DOC flux from the uplands (7.7 g C cm^{-2} y^{-1}; Table 4.4) is consistent with the estimated average annual flux from terrestrial ecosystems worldwide (Randerson et al., 2002). Therefore, the carbon flux from the uplands was equivalent to the worldwide average value, and the carbon flux from the forested wetlands and sloping bogs was substantially higher than the worldwide average. The substantial amount of carbon export makes the accounting of dissolved fluxes quite relevant in ecosystem carbon balance of the NCTR (Neff and Asner, 2001). The NCTR average watershed composition can be used to estimate the flux of DOC from a watershed with approximately 70% uplands, 15% forested wetlands, and 15% sloping bogs. This average estimated value for area weighted export from a total watershed (14.9 g C m^{-2} y^{-1}) compares favorably to the only published estimates of DOC flux from rivers in the region (Sugai and Burrell, 1982). The study of the Wilson and Blossom rivers by Sugai and Burrell (1984) estimated a DOC export of 13.3 g C m^{-2} y^{-1} from the low-gradient wetland dominated system in the Wilson River drainage and 7.9 g C m^{-2} y^{-1} in the upland dominated Blossom River Drainage.

4.5 Conclusions

Our results provide evidence that soil geomorphic assemblages provide a good approximation for functional attributes of DOC export from catchments in the NCTR. The overriding influence of shallow subsurface flow and the lack of groundwater interaction make wetland soil attributes effective predictors of catchment DOC flux. The timing of soil hydrologic fluctuations combined with seasonal temperature patterns influence the large quantities of DOC exported from the landscape in the late summer and fall.

4.6 References

Aitkenhead JA, McDowell WH (2000) Soil C:N ratio as a predictor of annual riverine DOC flux at local and global scales. Glob Biogeochem Cycles 14:127-138.

Aitkenhead-Peterson, JA, Alexander JE, Clair TA (2005) Dissolved organic carbon and dissolved organic nitrogen export from forested watersheds in Nova Scotia: identifying controlling factors. Glob Biogeochem Cycles 19:8.

Alexander EB, Kissinger E, Huecker RH, Cullen P (1989) Soils of southeast Alaska as sinks for organic carbon fixed from atmospheric carbon-dioxide. USDA Forest Service, Juneau, AK. Proceedings, Watershed '89: a conference on the stewardship of soil, air, and water resources.

Bengtson P, Bengtsson G (2007) Rapid turnover of DOC in temperate forests accounts for increased CO_2 production at elevated temperatures. Ecol Lett 10:783-790

Berg B, Berg M, Bottner P, Box E, Breymeyer A, Valvo de Anta R, Couteaux M, Escudero A, Gallardo A, Kratz W, Madeira W, Malkonen E, McClaugherty C, Meentenmeyer V, Munoz F, Piussi P, Remacle J, Vizr de Santo A (1993) Litter mass loss rates in pine forests of Europe and Eastern United States: some relationships with climate and litter quality. Biogeochemistry 20:127-159.

Berg B, McClaugherty C (2008) Plant litter: decomposition, humus formation, carbon sequestration. 2nd edn. Springer-Verlag, Berlin.

Bishop DM (1955) The relationship between precipitation and stream flow on a typical southeast Alaska stream. Technical Note 24, Alaska Forest Research Center, U.S. Department of Agriculture, Juneau, AK.

Bishop DM (1968) Interim report: hydrology of muskegs, Kadashan barometer watershed. U.S. Department of Agriculture, Forest Service, Juneau, AK.

Boyer E, Hornberger G, Bencala K, McKnight D (1997) Response characteristics of DOC flushing in an alpine catchment. Hydrological Processes 11:1635-1647.

Bridgham S, Johnston C, Pastor J, Updegraff K (1995) Potential feedbacks of northern wetlands on climate change. Bioscience 45:262-274.

Canham CD, Pace ML, Papaik MJ, Primack AGB, Roy KM, Maranger RJ, Curran RP, Spada DM (2004) A spatially explicit watershed-scale analysis of dissolved organic carbon in Adirondack lakes. Ecol Appl 14:839-854.

Caouette JP, DeGayner EJ (2005) Predictive mapping for tree sizes and densities in southeast Alaska. Landsc Urban Plan 72:49-63.

Clair TA, Pollock TL, Ehrman JM (1994) Exports of carbon and nitrogen from river basins in Canada's Atlantic Provinces. Glob Biogeochem Cycles 8:441-450.

Creed IF, Beall FD, Clair TA, Dillon PJ, Hesslein RH (2008) Predicting export of dissolved organic carbon from forested catchments in glaciated landscapes with shallow soils. Glob Biogeochem Cycles 22, GB4024, doi:10.1029/2008GB003294.

Dahl TE (2006) Status and Trends of Wetlands in the Conterminous United States 1998 to 2004. U.S. Department of the Interior; Fish and Wildlife Service, Washington, D.C. 112 pp.

Dalva M, Moore TR (1991) Sources and sinks of dissolved organic carbon in a forested swamp catchment. Biogeochemistry 15:1-19.

D'Amore DV, Ping CL, Herendeen PA (in preparation) Hydromorphic soil development in catenas of the Alaskan coastal temperate rainforest.

D'Amore DV, Edwards RT, Biles FE (submitted) Landscape controls on dissolved organic carbon concentrations of Alaskan coastal temperate rainforest streams. J Geophys Res.

D'Amore DV, Fellman JB, Edwards RT, Hood E (2010) Controls on dissolved organic matter concentrations in soils and streams from a forested wetland and sloping bog in southeast Alaska. Ecohydrol 3:249-261.

D'Amore DV, Lynn WC (2002) Classification of forested Histosols in Southeast Alaska. Soil Sci Soc Am J 66:554-562.

Dillon PJ, Molot L (1997) Effect of landscape form on export of dissolved organic carbon, iron and phosphorous from forested stream catchments. Water Resourc Res 33:2591-2600.

Driscoll CT, Fuller RD, Simone DM (1988) Longitudinal variations in trace metal concentrations in a Northern forested ecosystem. J Environ Qual 17:101-107.

Eckhardt B, Moore TR (1990) Controls on dissolved organic carbon concentrations in streams, southern Quebec. Can J Fish Aquat Sci 47:1537-1544.

Emili L, Price J (2006) Hydrological processes controlling ground and surface water flow from a hypermaritime forest-peatland complex, Diana Lake Provincial Park, British Columbia, Canada. Hydrol Proc 20:2819-2837.

Evans MG, Burt TP, Holden J, Adamson JK (1999) Runoff generation and water table fluctuations in blanket peat: evidence from UK data spanning the dry summer of 1995. J Hydrol 221:141-160.

Evans CD, Chapman PJ, Clark JM, Monteith DT, Cresser MS (2006) Alternative explanations for rising dissolved organic carbon export from organic soils. Glob Change Biol 12:2044-2053.

Fellman JB, D'Amore DV, Hood E, Boone RD (2008) Fluorescence characteristics and biodegradability of dissolved organic matter in forest and wetland soils from coastal temperate watersheds in southeast Alaska. Biogeochemistry 88:169-184.

Fellman JB, Hood E, Edwards RT, D'Amore DV (2009) Changes in the concentration, biodegradability and fluorescent properties of dissolved organic matter during stormflows in coastal temperate watersheds. J Geophys Research- Biogeosciences 114:G01021.

Fiedler S, Kalbitz K (2003) Concentrations and properties of dissolved organic matter in forest soils as affected by the redox regime. Soil Sci 168:793-801.

Findlay SM, Pace L, Lints D, Howe K (1991) Bacterial metabolism of organic carbon in the tidal freshwater Hudson Estuary. Mar Ecol Prog Ser 89:147-153.

Fraser CJD, Roulet, NT, Moore, TR (2001) Hydrology and dissolved organic carbon biogeochemistry in an ombrotrophic bog. Hydrol Proc 15:3151-3166.

Frey KE, Smith LC (2005) Amplified carbon release from vast West Siberian peatlands by 2100. Geophys Res Lett 32:1-4.

Freeman C, Evans CD, Monteith DT, Reynolds B, Fenner N (2001) Export of organic carbon from peat soils. Nature 412:785-785.

Gorham E 1991. Northern peatlands: Role in the carbon cycle and probable responses for global warming. Ecol Appl 1:182-195.

Gorham E, Underwood J, Janssens J, Freedman B, Maass W, Waller D, Ogden JI (1998) The chemistry of streams in southeastern and central Nova Scotia, with particular reference to catchment vegetation and the influence of dissolved organic carbon primarily from wetlands. Wetlands 18:115-132.

Gorham E, Lehman C, Dyke A, Janssens J, Dyke L (2007), Temporal and spatial aspects of peatland initiation following deglaciation in North America. Quart Sci Rev 26:300-311.

Harrison AF, Howard PJA, Howard DM, Howard DC, Hornung M (1995) Carbon storage in forest soils. Forestry 68:335-348.

Heusser C (1952) Pollen profiles from southeastern Alaska. Ecol Monogr 22:331-352.

Heusser C (1960) Late Pleistocene environments of North Pacific North America 35. American Geographical Society.

Holden J, Burt TP (2003) Hydrological studies on blanket peat: the significance of the acrotelm-catotelm model. J Ecol 91:86-102.

Hope D, Billett MF, Cresser MS (1994) A review of the export of carbon in river water: fluxes and processes. Environ Pollut 84:301-324.

Hope D, Billett MF, Milne R, Brown TAW (1997) Exports of organic carbon in British rivers. Hydrol Proc 11:325-344.

Hornberger GM, Bencala KE, McKnight DM (1994) Hydrological controls on dissolved organic carbon during snow melt in the Snake River near Montezuma, Colorado. Biogeochemistry 25:147-165.

Huntington TG, Aiken GR (2005). Dissolved organic carbon export from the Penobscot River Basin to the Gulf of Maine. American Geophysical Union Fall Meeting, San Francisco, CA. EOS Transactions of the American Geophysical Union 86 (52), abstract #B53A-07.

Kaiser K, Zech W (1998) Rates of dissolved organic matter release and sorption in forest soils. Soil Sci 163:714-725.

Kalbitz K, Solinger S, Park JH, Michalzik B, Matzner E (2000) Controls on the dynamics of dissolved organic matter in soils: a review. Soil Sci 165:277-304.

Leighty WW, Hamburg SP, Caouette J (2006) Effects of management on carbon sequestration in forest biomass in Southeast Alaska. Ecosystems 9:1051-1065.

Lin H (2003) Hydropedology: bridging disciplines, scales, and data. Vadose Zone J 2:1-11.

Lin H, Bouma J, Pachepsky Y, Western A, Thompson J, van Genuchten R, Vogel H J, Lilly A (2006) Hydropedology: synergistic integration of pedology and hydrology. Water Resour Res 42:W05301, doi:10.1029/2005WR004085.

Lloyd J, Taylor JA (1994) On the temperature dependence of soil respiration. Func Ecol 8:315-323.

Lumsdon DG, Stutter MI, Cooper RJ, Manson JR (2005) Model assessment of biogeochemical controls on dissolved organic carbon partitioning in an acid organic soil. Environ Sci Technol 39:8057-8063.

Lynn W, McKinzie W, Grossman R (1974) Field laboratory tests for characterization of Histosols, In: Aandahl A (ed) Histosols: their characteristics, classification, and use. Soil Sci Soc Amer, Madison, WI.

McDowell WH, Wood T (1984) Podzolization: Soil processes control dissolved organic carbon concentrations in streamwater. Soil Sci 137:23-32.

McDowell WH, Likens GE (1988) Origin, Composition, and Flux of Dissolved Organic Carbon in the Hubbard Brook Valley. Ecol Monogr 58:177-195.

Meybeck M (1982) Carbon, nitrogen and phosphorus transport in world rivers. Amer J Sci 282:401-450.

Michalzik B, Kalbitz K, Park JH, Solinger S, Matzner E (2001) Fluxes and concentrations of dissolved organic carbon and nitrogen – a synthesis for temperate forests. Biogeochemistry 52:173-205.

Miller RD (1973) Gastineau Channel formation: a composite glaciomarine deposit near Juneau, Alaska. Geological Survey Bulletin, Contributions to stratigraphy, USGS Report #1394, Washington, DC.

Mitchell C, Branfireun B (2005) Hydrogeomorphic controls on reduction–oxidation conditions across boreal upland–peatland interfaces. Ecosystems 8:731-747.

Moore TR (1989) Dynamics of dissolved organic carbon in forested and disturbed catchments, Westland, New Zealand: 1. Maimai. Water Resour Res 25:1321-1330.

Moore TR, RJ Jackson (1989) Dynamics of dissolved organic carbon in forested and disturbed catchments, Westland, New Zealand: 2. Larry River. Water Resour Res 25:1330-1341.

Moore TR, De Souza W, Koprivnjak JF (1992) Controls on the sorption of dissolved organic carbon by soils. Soil Sci 154:120-129.

Moore T, Blodau C, Turunen J, Roulet N, Richard, PJ (2004) Patterns of nitrogen and sulfur accumulation and retention in ombrotrophic bogs, eastern Canada. Glob Change Biol 11:356-367.

Mulholland PJ (2003) Sources, production, and regulation of allochthonous dissolved organic matter inputs to surface waters. In: Findlay SEG, Sinsabaugh RL (eds) Aquatic ecosystems: interactivity of dissolved organic matter. Elsevier, New York, pp 25-70.

Mulholland P, Kuenzler E (1979) Organic carbon export from upland and forested wetland watersheds. Limnol Oceanogr 24:960-966.

Nadelhoffer K, Giblin A, Shaver G, Laundre J (1991) Effects of temperature and substrate quality on element mineralization in six arctic soils. Ecology 72:242-253.

Neiland BJ (1971) The forest-bog complex in Southeast Alaska. Vegetatio 22:1-63.

Neff JC, Asner GP (2001) Dissolved organic carbon in terrestrial ecosystems: synthesis and a model. Ecosystems 4:29-48.

Neff JC, Chapin FS, Vitousek PM (2003) Breaks in the cycle: dissolved organic nitrogen in terrestrial ecosystems. Front Ecol Environ 1:205-211.

Nelson PN, Baldock JA, Oades JM (1993) Concentrations and composition of dissolved organic carbon in streams in relation to catchment soil properties. Biogeochemistry 19:27-50.

Nowacki G, Krosse P, Fisher G, Brew D, Brock T, Shephard M, Pawuk W, Baichtal J, Kissinger E (2001) Ecological subsections of southeast Alaska and neighboring areas of Canada. USDA Forest Service, Alaska Region, Technical Publication R10-TP-75 (2001).

Pastor J, Solin J, Bridgham SD, Updegraff K, Harth P, Weishampel P, Dewey B (2003) Global warming and the export of dissolved organic carbon from boreal peatlands. Oikos 100:380-386.

Qualls RG, Haines BL, Swank WT (1991) Fluxes of dissolved organic nutrients and humic substances in a deciduous forest. Ecology 72:254-266.

Qualls RG, Haines B (1992) Biodegradability of dissolved organic matter in forest throughfall, soil solution, and stream water. Soil Sci Soc Am J 56:578-586.

Qualls R.(2000) Comparison of the behavior of soluble organic and inorganic nutrients in forest soils. For Ecol Manage 138:29-50.

Randerson JT, Chapin III FS, Harden JW, Neff JC, ME Harmon (2002) Net ecosystem production: a comprehensive measure of net carbon accumulation by ecosystems. Ecol Appl 12: 937-947.

Raymond PA, Saiers JE (2010) Event controlled DOC export from forested watersheds. Biogeochemistry doi10.1007/s10533-010-9416-7.

Runkel RL, Crawford CG, Cohn TA (2004) Load estimator (LOADEST): a FORTRAN program for estimating constituent loads in streams and rivers: U.S. Geological Survey Techniques and Methods Book 4, Chapter A5, 69 pp.

SAS Institute (2003) Version 9.2, SAS Institute Inc., Cary, NC, USA.

Schiff SL, Aravena R, Trumbore SE, Hinton MJ (1997) Export of DOC from forested catchments on the Precambrian Shield of Central Ontario: clues from ^{13}C and ^{14}C. Biogeochemistry 36:43-65.

Schimel D, Braswell B, Holland E, McKeown E, Ojima D, Painter T, Parton W, Townsend A (1994) Climatic, edaphic and biotic controls over storage and turnover of carbon in soils. Glob Biogeochem Cycles 8:279-293.

Soil Survey Division Staff. 1999. Soil Survey Manual. USDA Natural Resources Conservation Service, Government Printing Office, Washington, DC.

Sugai SF, Burrell DC (1984) Transport of dissolved organic carbon, nutrients, and trace metals from the Wilson and Blossom rivers to Smeaton Bay, southeast Alaska. Can J Fish Aq Sci 41:180-190.

Turner DP, Koerper GJ, Harmon ME, Lee JJ (1995) A carbon budget for forests of the conterminous United States. Ecol Appl 5:421-436.

USDA Forest Service (1997) Tongass National Forest Land and Resource Management Plan. Juneau, USDA Forest Service Region 10.

USFWS (2009) Classification of wetlands and deepwater habitats of the United States. National Wetlands Inventory website. US Dept Interior; Fish and Wildlife Service, Washington D.C. http://www.fws.gov/wetlands.

Worrall F, Burt T, Adamson J (2004) Can climate change explain increases in DOC flux from upland peat catchments? Sci Tot Environ 326:95-112.

Worrall F, Burt TP (2008) The effect of severe drought on the dissolved organic carbon (DOC) concentration and flux from British rivers. J Hydrol 361:262-274.

Xenopoulos MA, Lodge DM, Frentress J, Kreps TA, Bridgham SD, Grossman E, Jackson CJ (2003) Regional comparisons of watershed determinants of dissolved organic carbon in temperate lakes from the Upper Great Lakes region and selected regions globally. Limnol Oceanogr 48:2321-2334.

Yano Y, McDowell WH, Aber J (2000) Biodegradable dissolved organic carbon in forest soil solution and effects of chronic nitrogen deposition. Soil Biol Biochem 32:1743-1751.

Zheng D, Hunt E Jr, Running SW (1993) A daily soil temperature model based on air temperature and precipitation for continental applications. Clim Res 2:183-191.

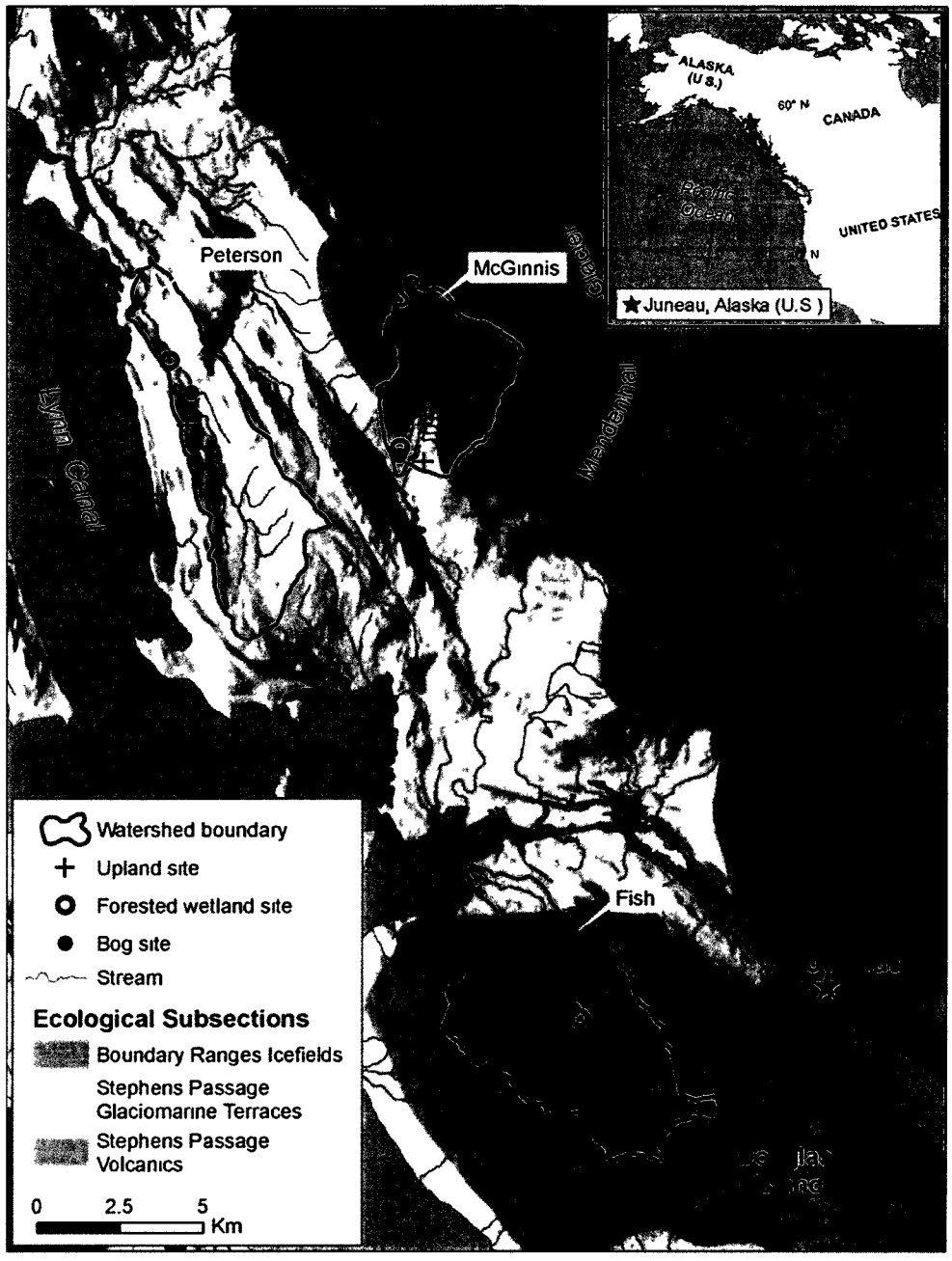

Figure 4.1. The ecological subsections and watersheds of the Juneau, AK area. The approximate locations for measurements of dissolved organic carbon flux in sloping bogs, forested wetlands, and uplands are identified within the watershed (Peterson, McGinnis, and Eaglecrest/Fish) boundaries.

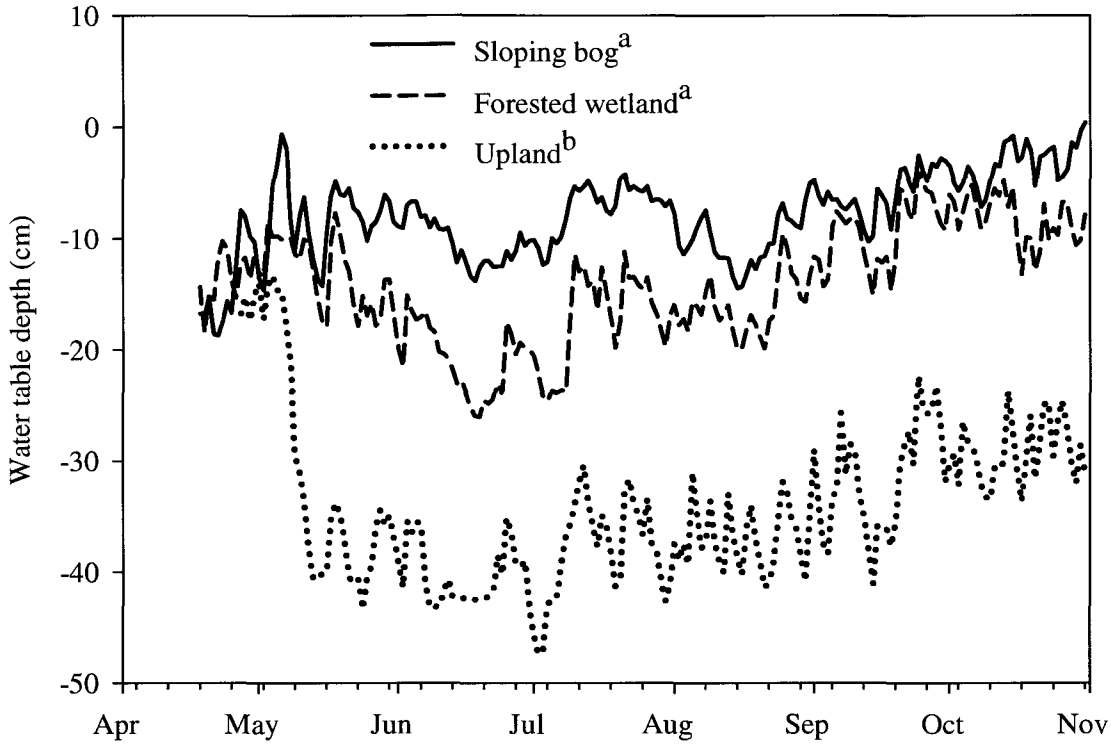

Figure 4.2. Soil water table fluctuations in sloping bogs, forested wetlands, and uplands. Data from 2006 and 2007 are combined. Letters next to site legends represent significant differences in average water table depth at alpha = 0.50; model ($F_{(2,112)} = 102.75$, $P < 0.001$).

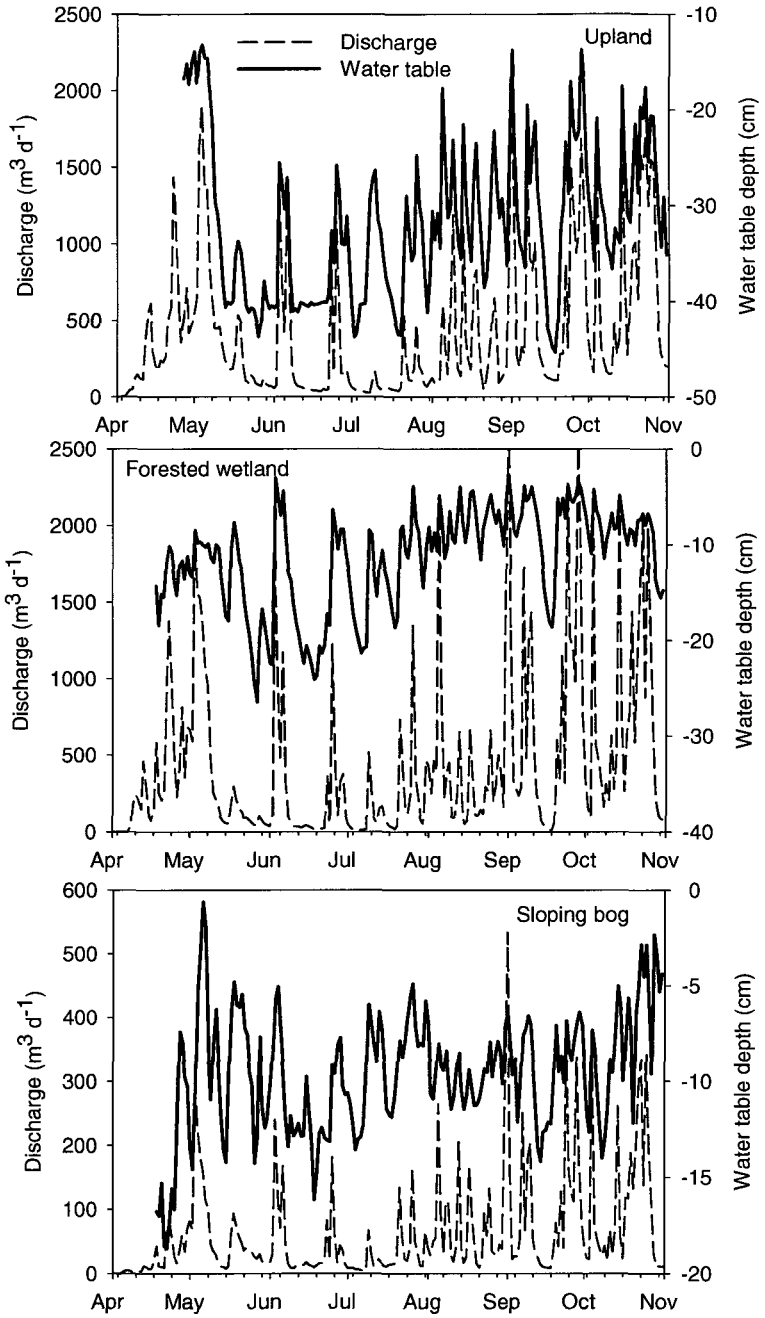

Figure 4.3. Water table and streamwater fluctuations in sloping bogs, forested wetlands, and uplands. Streamwater discharge and water table depth are significantly correlated in all sites: Bog $r = 0.47$; $P < 0.001$; Forested wetland $r = 0.61$; $P < 0.001$; Upland $r = 0.81$; $P < 0.001$. Note scale changes for water table depth at all sites and discharge in the sloping bog.

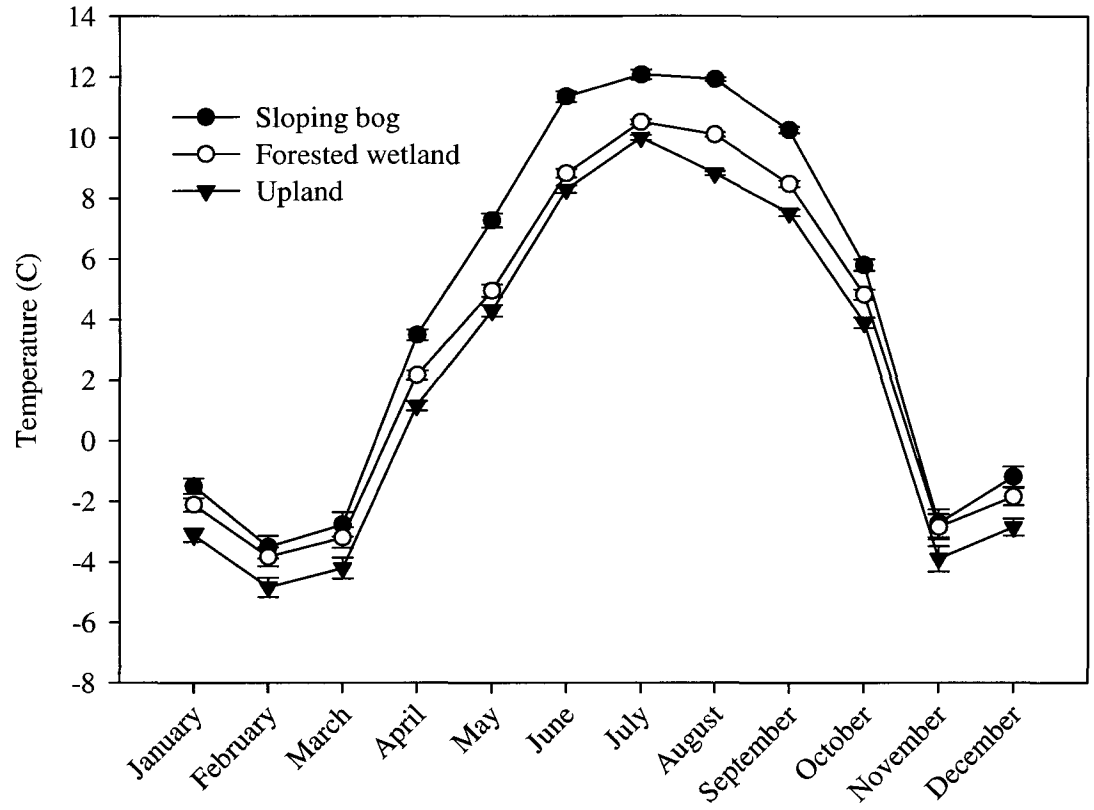

Figure 4.4. Average monthly soil temperatures in sloping bogs, forested wetlands, and uplands at 10-cm depth. Values are mean ± SE, n = 3 per hydropedological unit type.

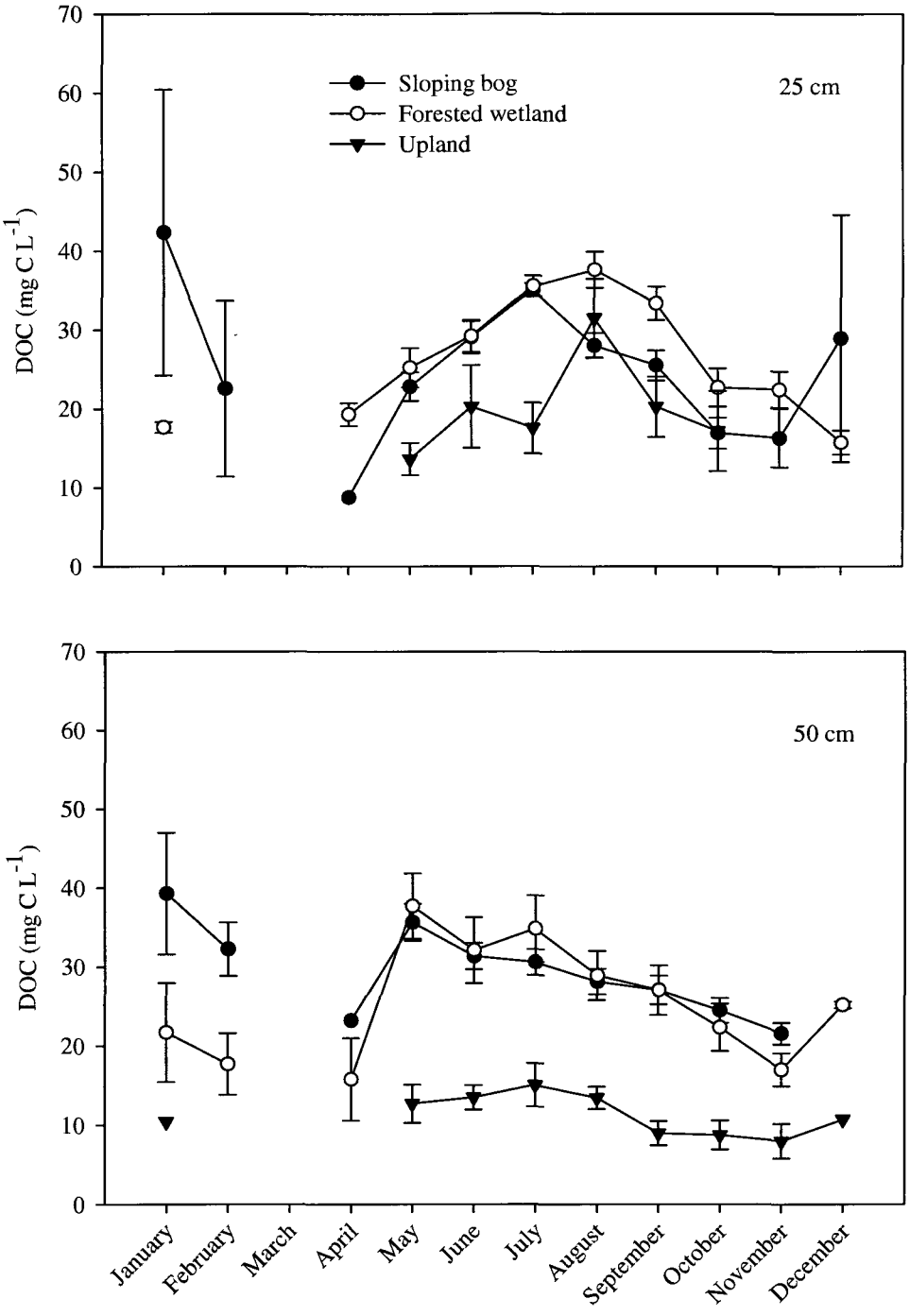

Figure 4.5. Seasonal patterns of soil dissolved organic carbon concentrations in sloping bogs, forested wetlands, and uplands. Measurements were taken at 25- and 50-cm depths. The soil solution concentrations are means ± SE (n = 3 per hydropedologic type).

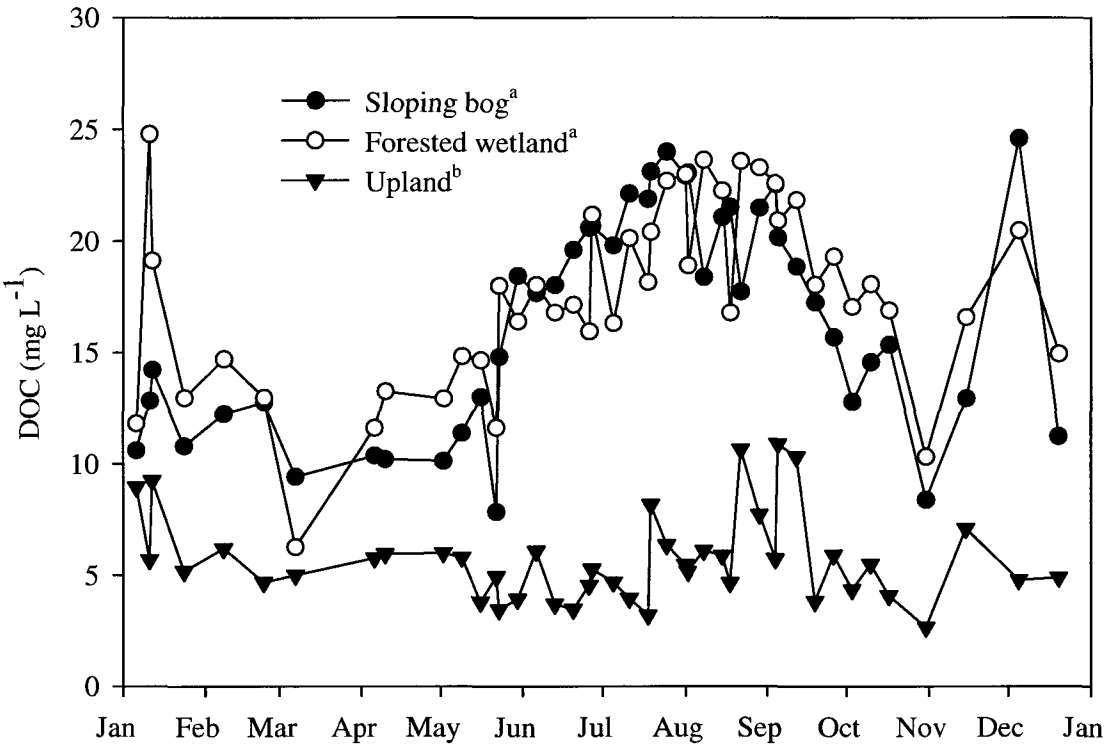

Figure 4.6. Seasonal patterns of streamwater dissolved organic carbon concentrations in sloping bogs, forested wetlands, and uplands. Data from 2006 and 2007 are combined, n = 3 for each hydropedologic type. Letters next to site legends represent significant differences in average DOC concentration among sites at alpha = 0.50; model ($F_{(2,112)} = 91.32$, $P < 0.001$).

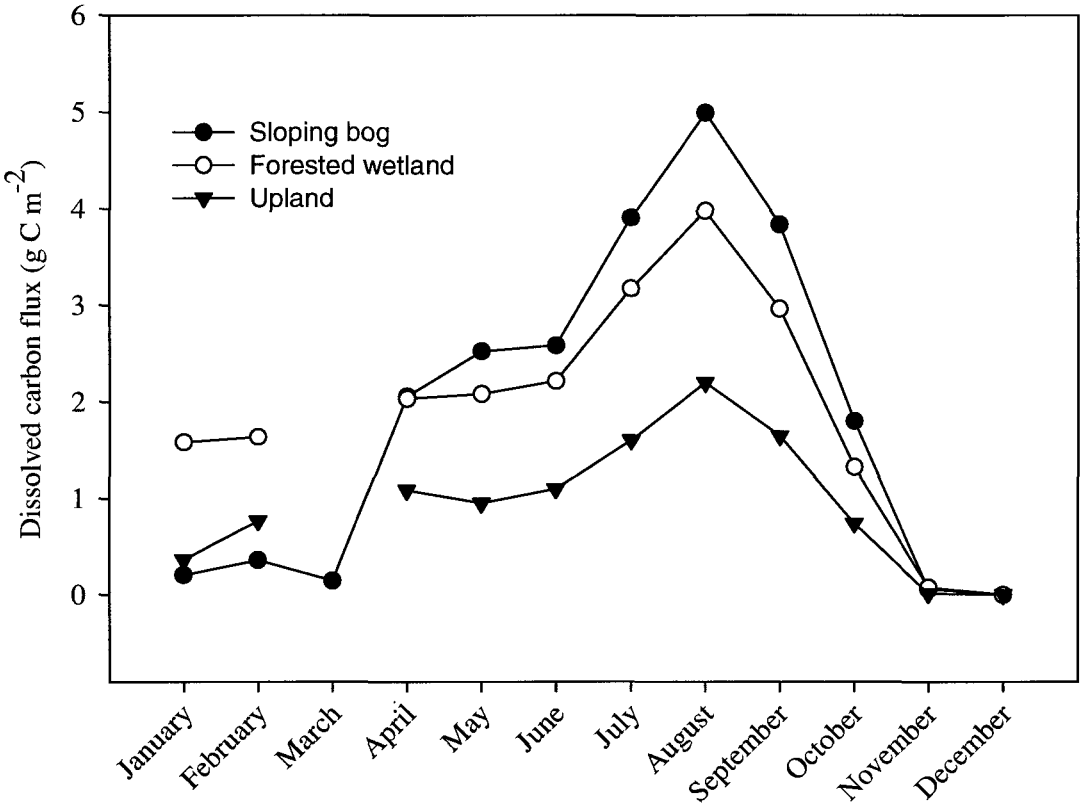

Figure 4.7. Monthly dissolved organic carbon flux derived from average fluxes among replicate values for sloping bogs, forested wetlands, and uplands.

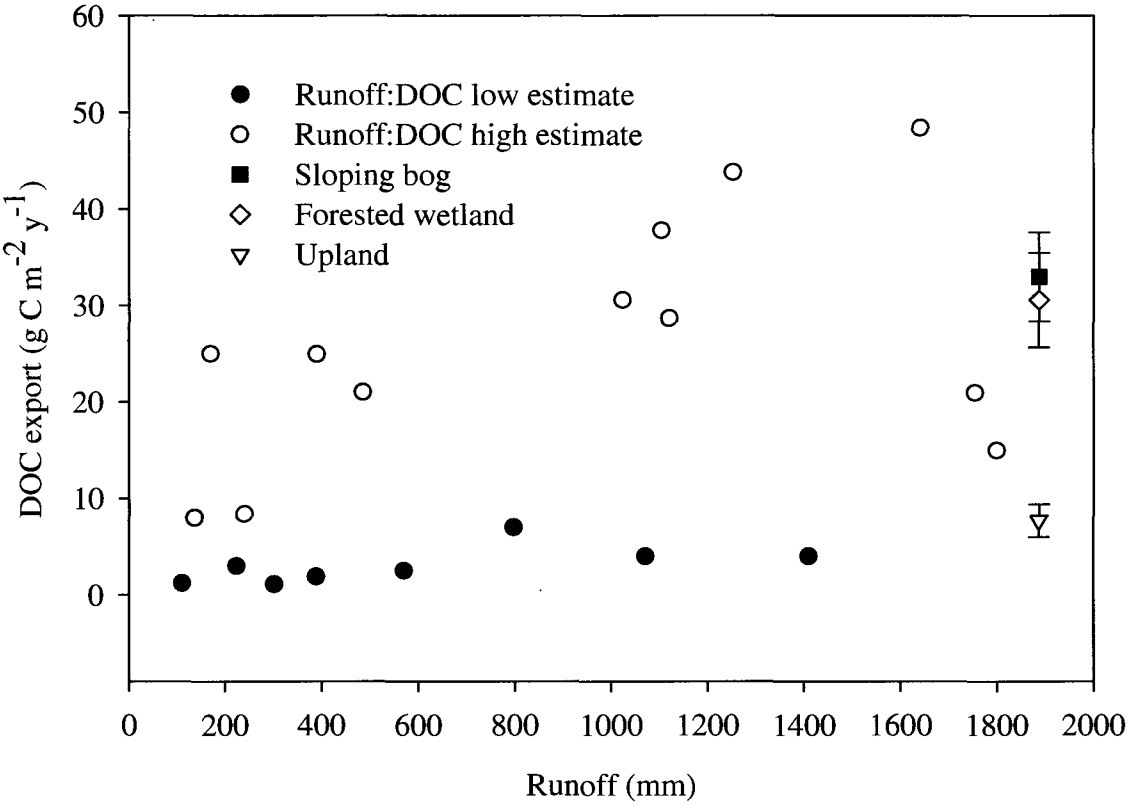

Figure 4.8. Relationship between streamwater runoff and disolved organic carbon export. Data for the amount of runoff and dissolved organic carbon (DOC) (from Fraser et al., 2001) were divided into low (<5) and high (>5) DOC export for comparison with sloping bogs, forested wetlands, and uplands in this study.

Table 4.1.Soil classification and attributes of hydropedologic types within three watersheds in the Juneau, AK area.

Watershed	Hydropedological type	Site code	Taxonomic classification[†]	Elevation (m)	Slope (%)	Drainage class[§]
Eaglecrest	Sloping bog	EBT	Dysic Typic Cryohemist	89	3	Very poorly-drained
	Forested Wetland	EFW	Coarse-loamy, mixed, dysic Terric Cryohemists	248	5	Poorly-drained
	Upland	EUT	Coarse-loamy, mixed, superactive, Typic Haplocryod	392	15	Well-drained
McGinnis	Sloping bog	MBT	Dysic Typic Cryohemist	133	3	Very poorly-drained
	Forested Wetland	MFW	Loamy, mixed, Terric Cryohemists	128	19	Somewhat poorly-drained
	Upland	MUT	Coarse-loamy, mixed, superactive, Typic Humicryod	202	20	Well-drained
Peterson	Sloping bog	PBT	Dysic Typic Cryohemist	112	2	Very poorly-drained
	Forested Wetland	PFW	Coarse-loamy, mixed, Histic Cryaquept	20	5	Poorly-drained
	Upland	PUT	Loamy, mixed, superactive, Lithic Haplocryod	163	20	Well-drained

[†]Taxonomic classification represents the field and laboratory interpretation for the site, not the official series.

[§]Drainage class was determined by field observations and soil map unit assignment from soil resource inventory.

Table 4.2. Physical properties of soils studied at three replicated sites in southeast Alaska.

Site	Depth (cm)	Horizon	Matrix color	Bulk Density (Mg m^{-3})	Fiber content (%)		Pyrophosphate Color	Percent			kg C m^2		Texture
					Unrubbed	Rubbed		C	N	C:N	C	N	
EBT	0-13	Oi	10YR 3/4	0.08	74	60	10YR 8/3	45.3	2.08	21.78	35.16	1.61	Fi
	13-47	Oe1	10YR 3/3	0.11	42	22	10YR 8/2	48.5	2.34	20.76	51.05	2.46	He
	47-84	Oe2	5YR 3/2	0.12	54	24	10YR 8/3	52.7	1.86	28.37	64.87	2.29	He
	84-130+	Oe3	5YR 2.5/2	0.15	68	20	10YR 8/3	48.5	2.06	23.55	71.41	3.03	He
EFW	0-4	Oi	5YR 5/6	0.07	82	66	10YR 8/3	46.5	0.82	56.71	31.87	0.56	Fi
	4-12	Oe1	5YR 3/2	0.10	72	24	10YR 8/3	47.5	2.54	18.71	49.29	2.63	He
	12-25	Oe2	10YR 2/2	0.13	84	44	10YR 7/4	49.9	2.20	22.68	64.94	2.86	He
	25-43	Oe3	10YR 2/2	0.15	68	12	10YR 4/3	50.0	1.94	25.80	73.41	2.85	He
	43-71	B	2.5Y 3/2	0.38				8.28	0.19	43.58	31.85	0.73	L
	71-91	Bc	5Y 5/3	0.64				nd	0.03	nd	3.20	0.00	SL
	91+	C	5GY 6/1	1.15				0.77	0.00	nd	8.83	0.00	SL

Table 4.2. continued

EUT	0-9	Oi	7.5YR 2.5/3	0.05	72	40	10YR 8/3	53.32	1.01	52.79	28.58	0.54	Fi
	9-12	Oe	5YR 2.5/2	0.09	52	32	10YR 8/4	49.8	0.87	57.25	42.37	0.74	He
	12-14	E	7.5YR 5/2	0.42				6.33	0.21	30.14	26.85	0.89	SiL
	14-20	Bh	5YR 3/3	0.24				9.77	0.39	25.05	23.09	0.92	SL
	20-48	Bs	5YR 4/6	0.31				5.30	0.16	33.13	16.57	0.50	SL
	48-72	BC	2.5Y 4/3	0.48				2.09	0.08	26.13	10.12	0.39	L
	72+	C	2.5Y 5/4	0.84				0.0	0.01	nd	4.20	0.00	L
MBT	3-12	Oi2	10YR 3/4	0.05	84	76	10YR 7/2	49.1	1.08	45.47	26.58	0.58	Fi
	12-23	Oe1	10YR 3/6	0.08	71	62	10YR 8/1	47.7	1.29	37.04	37.97	1.03	He
	23-35	Oe2	10YR 5/6	0.08	66	44	10YR 8/1	49.5	1.44	34.38	39.71	1.15	He
	35-45	Oe3	5YR 3/1	0.08				45.0	1.40	45.00	36.00	0.80	
	45-84	Oe4	5YR 4/3	0.20	64	60		48.0	1.00	48.00	96.00	2.00	Fi
	84-132	Oe5	5YR 3/2	0.21				53.0	8.00	6.63	111.30	16.80	He
	132-200	Oe6	5YR 4/6	0.26				37.0	5.00	7.40	96.20	13.00	He

Table 4.2. continued

MFW	0-4	Oi	5YR 3/1	0.20				53.00	1.30	40.77	106.00	2.60	Fi
	4-23	Oa	5YR 4/3	0.21				48.00	1.10	43.64	100.80	2.31	Sa
	23-45	Oe	7.5YR 3/3	0.25				57.00	1.70	33.53	142.50	4.25	He
	45-70	Oe	7.5YR 4/4	0.24				42.00	0.80	52.50	99.20	1.92	He
	70-87	B	5YR 3/2	0.62				16.00	0.50	32.00	99.20	3.10	SL
	87-108	Bg	4YR 10/3	1.50				7.00	0.40	17.50	105.00	6.00	SiL
	108-162	Cg	2.5YR 5/3	1.30				nd	nd	nd	nd	nd	SiL
MUT	0-10	Oi	2.5YR 2.5/3	0.10	84	16	10YR 8/2	53.3	0.86	62.07	55.80	0.90	Fi
	10-17	Oe	2.5YR 2.5/2	0.14	68	4	7.5YR 4/3	24.0	1.04	23.09	34.75	1.51	He
	17-29	O/E	7.5YR 2.5/2	0.43	66	10	10YR 5/4	12.6	0.63	20.06	55.94	2.74	
	29-43	Bh	5YR 3/3	0.19				8.50	0.30	28.33	16.17	0.57	L
	43-65	Bs	7.5YR 4/6	0.25				4.99	0.18	27.72	12.33	0.44	SL
	65-85	B/C	10YR 4/4	nd				1.78	0.09	19.78	nd	nd	LS
	85+	C	10YR 4/3	nd				2.54	0.13	19.54		nd	LS

Table 4.2. continued

PBT	0-9	Oe1	10YR 3/4	0.13	56	36	10YR 8/2	50.3	1.82	27.68	66.09	2.39	He
	9-18	Oe2	10YR 3/6	0.10	60	24	10YR 7/2	52.3	1.32	39.67	53.35	1.34	He
	18-30	Oe3	10YR 3/3	0.10	54	30	10YR 7/2	52.4	1.29	40.64	54.53	1.34	He
	30-38	Oe4	10YR ¾	0.17	82	20	10YR 6/3	55.4	1.15	48.25	92.93	1.93	He
	38-69	Oi	10YR 3/6	0.1	64	30	2.5Y 7/2	52.6	1.03	51.10	53.61	1.05	Fi
	69+	Oe5	10YR 3/4	0.12	52	14	2.5Y 8/2	56.5	1.62	34.90	70.56	2.02	He
PFW	0-5	Oa1	7.5YR 2.5/1	0.25	66	16	10YR 4/4	27.9	1.28	21.84	69.49	3.18	He (Sa)
	5-26	Oa2	10YR 3/2	0.39	40	0	10YR 4/3	13.5	0.68	19.87	52.17	2.63	Sa
	26-40	Bw1	10YR 2/2	0.33	44	10	10YR 4/4	6.45	0.25	25.80	21.33	0.83	SL
	40-68	Bw2	10YR 2/2	0.50				2.51	0.08	31.38	12.56	0.40	SL
	68+	C	2.5YR 3/2	0.77				0.51	0.00	nd	3.95	0.00	LS
PUT	0-4	Oe1		0.1	84	68	10YR 8/2	51.0	1.04	49.09	50.25	1.02	He
	4-10	Oe2		0.08	88	68	10YR 8/2	50.5	0.89	56.75	40.41	0.71	He
	10-18	E		nd				13.5	0.67	20.15	Nd	Nd	L
	18-26	Bh		0.38				8.43	0.29	29.07	32.41	1.11	L
	26-54	Bs		0.63				3.55	0.10	35.50	22.51	0.63	SiL
	54+	Cr		nd				nd	nd	nd	nd	nd	R

Table 4.3. Parameter estimates for temperature dependent dissolved organic carbon models in soils of sloping bogs, forested wetlands, and uplands at depths of 25 and 50 cm. Parameters estimates Parameter estimates (± SD) for activation energy (E_0 [K^{-1}]; Q_{10}) and flux at 10° C ($R_{ref.}$) were fit using Lloyd and Taylor (1994) and Q_{10} models.

Hydropedologic type	Depth (cm)	Lloyd & Taylor model			Q_{10} model		
		E_0	R_{ref}	P	Q_{10}	R_{ref}	P
Sloping bog	25	329.27 ± 62.15	25.82 ± 1.23	<0.001	3.03 ± 0.66	25.46 ± 1.28	<0.001
	50	60.27 ± 53.10	29.31 ± 1.66	<0.001	1.22 ± 0.22	29.21 ± 1.66	<0.001
Forested wetland	25	212.96 ± 29.73	35.17 ± 1.13	<0.001	2.15 ± 0.23	35.26 ± 1.12	<0.001
	50	136.01 ± 91.70	32.29 ± 3.59	<0.001	1.62 ± 0.54	32.28 ± 3.66	<0.001
Upland	25	177.65 ± 157.60	23.71 ± 4.01	0.003	1.86 ± 1.05	23.75 ± 4.14	0.003
	50	160.02 ± 100.90	14.00 ± 1.55	<0.001	1.80 ± 0.64	14.12 ± 0.64	<0.001

Table 4.4. Mean concentration of dissolved organic carbon (DOC) and DOC flux in sloping bogs, forested wetlands, and uplands in the Juneau, AK area.

Watershed Hydropedologic type	Area (ha)	DOC ($mg\ L^{-1}$)	Annual DOC export ($kg\ ha^{-1}$)	($kg\ ha^{-1}\ mm^{-1}$)	Annual area specific DOC export ($g\ C\ m^{-2}\ y^{-1}$)
Peterson					
Sloping bog	1	17.49	244.93	0.17	24.49
Forested wetland	5	15.22	233.54	0.17	23.35
Upland	20	6.62	82.00	0.06	8.20
McGinnis					
Sloping bog	0.5	22.80	341.98	0.18	34.20
Forested wetland	5	27.12	284.72	0.18	28.47
Upland	1	3.68	105.17	0.06	10.52
Eaglecrest					
Sloping bog	1	18.08	401.57	0.18	40.16
Forested wetland	10	24.64	400.48	0.18	40.05
Upland	8	13.87	45.15	0.06	4.52
All sloping bog	0.8 (0.2)		329.5 (45.6)		33.0 (4.6)
All forested wetland	6.7 (1.7)		306.2 (49.4)		30.6 (4.9)
All upland	9.7 (5.5)		77.4 (17.5)		7.7 (1.7)

Chapter 5: Terrestrial and stream chemical linkages in Alaskan coastal temperate rainforest catchments[1]

5.1 Abstract

An important aspect of watershed research is understanding terrestrial ecosystem development within the context of biogeochemical evolution. The working hypothesis for forested ecosystems is the accumulation and conservation of elements derived from weathering and nitrogen fixation early in succession, and increasing losses during later stages of development. The North American perhumid coastal temperate rainforest (NCTR) provides an ideal natural setting to conduct watershed mass-balance and biogeochemical studies. The post-glacial landscapes of the NCTR provide an array of ecosystem types that provide building blocks for compiling estimates of watershed input-output budgets. In this study, we determined the geochemical signature of watersheds and catchments across a hydrologic gradient in the NCTR to provide baseline data for this remote region. The mainstem watersheds and upland catchments had weathering signatures where Ca^{2+} and Mg^{2+} were exported in excess of predicted atmospheric inputs to the areas. Biological production in the poorly-drained wetlands diminishes the weathering signature and export of inorganic carbon. Input-output budgets quantified the HCO_3^- and organic acid alkalinity through estimates of charge balance in streamwaters draining individual sub-catchments distinguished by soil hydropedological units. The potential for inorganic carbon export and movement of free CO_2 makes the soil-stream linkage compelling for overall ecosystem carbon cycling.

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[1] D'Amore, DV, Trainor TP, Edwards RT, Valentine DW (to be submitted to Geoderma) Terrestrial and stream chemical linkages in Alaskan coastal temperate rainforest catchments.

5.2 Introduction

An important aspect of watershed research is understanding terrestrial ecosystem development within the context of biogeochemical evolution. The working hypothesis for forested ecosystems is the accumulation and conservation of elements derived from weathering and nitrogen fixation early in succession, and increasing losses during later stages of development (Vitousek and Reiners, 1975). Therefore, an assessment of the relative state of ecosystem development can be obtained through an analysis of geochemical balances in catchments or watersheds. A critical question underlying these studies is the prevalence of weathering, or the interaction of precipitation with underlying parent material. Studies of terrestrial ecosystem development have relied on soil and regolith diversity to evaluate how internal and external element supplies can reveal historical and present terrestrial ecological patterns (Hedin and Hetherington, 1995; Hedin et al., 1995). Soils are a key component in understanding how long-term weathering patterns and biogeochemical cycling are associated with environmental change (Birkeland, 1984; Richter et al., 2007). The use of soil state factors is a valuable tool for partitioning diverse landscapes into functional units for evaluation of chemical budgets (Perakis and Hedin, 2007).

Direct measurements of soil and rock weathering provide complementary information to watershed mass balance studies from stream geochemical analysis (Johnson et al., 1968; Johnson 1971; Price et al., 2005). Many watershed mass-balance and soil chemical evaluations were undertaken to address concerns regarding acid deposition on soils and waters once acid rain was recognized as an environmental threat in forests of the Northeastern U.S. and Europe. In the U.S., Reuss and Johnson's (1986) model has provided the key components for many applications, while in Europe, the profile model (Sverdrup and Warfvinge, 1988; 1991) has been applied in a similar manner to evaluate changes in watershed chemistry due to soil acidification. An emerging concern along coastal margins is the impact of varying acidic inputs and alkalinity of the river water (Salisbury et al., 2008). The prevalence of highly acidic river output in remote areas is a concern where there is not a rich history of watershed research, such as the Arctic and Northern Pacific coastal margin (Holmes et al., 2008). Over long time periods, these areas provide a biological feedback through weathering and watershed outputs that provides a mitigating impact on global carbon cycles through silicate weathering (Berner and Berner, 1987; Caldeira, 2006). However, the short-term (i.e. decadal) shifts in streamwater alkalinity may influence the

biological processes in estuaries in northern coastal marine ecosystems where organic acids carried from source pools in the terrestrial system can drastically impact the estuary chemistry over time.

Many of the fundamental concepts regarding watershed management have come from long-term watershed studies that integrate measurements of soils, vegetation, and water. The development of terrestrial ecosystems is heavily influenced by climate, and there is a broad representation in the literature among worldwide climates in the U.S. (Bormann and Likens, 1979; Ice and Stednick, 2004; Lugo et al., 2006), Europe (Christophersen et al., 1982), Hawaii (Kennedy et al., 1998), and South America (Hedin et al., 1995). The diversity of biomes worldwide provides templates for examining the influence of climate on the evolution of terrestrial ecosystems. Relatively pristine areas are valuable sites for testing theories and applying models of watershed scale ecosystem development (Crews et al., 1995; Hedin et al., 1995; Vitousek and Farrington, 1997). However, these pristine areas generally lack long-term stream biogeochemical data due to their remote settings. The post-glacial landscapes of the North American perhumid coastal temperate rainforest (NCTR; Alaback, 1996) have provided a template for a rich history of post-glacial chronosequence studies (Chandler, 1942; Chapin et al., 1994), yet there are few studies that provide watershed input-output budget estimates for the region (Stednick, 1981; Sugai and Burrell, 1984). The NCTR provides an ideal natural setting to conduct watershed mass-balance and biogeochemical studies to build models that address the concerns along coastal margins.

In this study we assessed the geochemical signature among replicated sub-catchments across a hydrologic gradient in the NCTR to provide baseline data for this remote region. Our goal was to address three key questions: 1) Is there a weathering signature among the catchments and mainstem reaches?; 2) Is the system a source, sink, or steady-state relative to element budgets?; and 3) How does the export of elements influence the charge balance and inorganic carbon fluxes across the hydropedologic gradient (Chapter 2).

5.3 Materials and methods

5.3.1 Site descriptions

Research was conducted in three watersheds near Juneau, AK that are part of a long-term carbon flux study in the North American Carbon Program (NACP; Figure 5.1). We chose watersheds in three different ecological subsections as the core sites to address the higher order

hydrogeomorphic control on ecosystem functions (Figure 5.1). The watersheds represent three distinct landscapes dominated by different lithology and dominant forms of landscape evolution (Nowacki et al., 2001). Peterson watershed, in the Stephens Passage glaciomarine terrace subsection, is primarily composed of slowly permeable glaciomarine sediments (Miller, 1973) along with bedrock outcrops that occur on moderate to low slopes. Peterson watershed represents an end-member type watershed dominated by wetlands (53% of watershed area). In contrast, McGinnis watershed is primarily composed of recently deglaciated areas within the Boundary Ranges Icefield subsection and has low wetland coverage (<5% of watershed area). Eaglecrest watershed is composed of intrusive volcanic and sedimentary rock in the Stephens Passage volcanic subsection and represents a mix of physiographic features that include alpine, productive forest, and lowland wetland ecosystems. We derived soil classes that were primarily identified by hydrologic function from available surveys and inventories and delineated three sites within each of these three watershed blocks. We define these individual sites as distinct hydropedologic functional types following the concept of hydropedology proposed by Lin (2003; also see Lin et al., 2006; Chapter 2, Table 2.1). The three hydropedologic types were stratified by drainage class which was derived from the soil map unit. Histosols and inceptisols were present in both the sloping bog and forested wetland sites, and the hydrologic control section was defined by the depth of the acrotelm boundary (D'Amore et al., 2010). In contrast, uplands were defined by a lack of near surface soil saturation and were located on higher gradient slopes with zones of deep soil development. The hydropedologic units are consistent with the wetland classification categories of sloping bog, forested wetland, and upland. Wetlands are roughly mapped by the National Wetland Inventory (NWI), but low resolution and inaccuracies dictate that field observations of soils patterns be used to more accurately delineate the boundaries of those classes in the field. Sub-catchments that were dominated by each hydropedologic unit were delineated and a flow-gauging weir was installed to measure chemical export from each site. The distinct hydropedologic units were hypothesized to control soil hydrology and influence the overall hydrologic and biogeochemical behavior of the sub-catchment (Table 5.1).

5.3.2 Climate

The Juneau area is characterized by mild temperatures, abundant cloud cover, and heavy precipitation. The climate of the study sites is similar to the weather station at Juneau airport, therefore, the records reported for the airport are used as reference conditions for the replicate

sub-catchments. The growing season at the airport averages 146 days from May 4 to September 28. February to June is the period of lightest precipitation with monthly averages of 7.6 cm. The heaviest precipitation is from June to October with averages of 15.6 to 17.8 cm per month. Snowfall occurs rarely in October at lower elevations while the majority of the snowfall occurs between November and March. Most low to middle elevations are free of snow by April. Mean annual precipitation is 140.9 cm, mean annual snowfall is 219.2 cm, and mean annual temperature is 4.7° C.

5.3.3 Soil and streamwater sampling and analysis

Soil profile characterization data was obtained from a representative soil pedon located within each sub-catchment (Soil Survey Division Staff, 1999; Table 5.1), which were then classified according to soil taxonomy (Soil Survey Staff, 2010). Atmospheric deposition data was obtained from the local National Atmospheric Deposition site (NADP) located in the city and borough of Juneau near the University of Alaska, southeast campus. Streamwater samples were taken from outlet streams draining each sub-catchment and mainstem outlet streams from each watershed. Streamwater samples were field-filtered using pre-combusted, Gelman A/E glass fiber filters and stored in the refrigerator until analysis. Stream water samples were analyzed for cations and anions by ICP and/or flame absorption spectrophotometry. Dissolved concentrations of Ca^{2+}, Mg^{2+}, Na^{+}, K^{+}, Si, Al^{3+}, and Fe^{3+} were determined on time series data from three seasonal samples (June, August, October) in 2007, and monthly samples taken from May-October during 2009. Si values are only available for 2007, so catchment and watershed interpretations for atmospheric and terrestrial input-output budgets were constructed with this data in order to compare stoichiometric weathering patterns from alumino-silicates in the watersheds. The terrestrial contribution of cations was determined by calculating an enrichment factor for each element from atmospheric inputs. Predicted element to Cl^{-} relationships in streamwater were calculated from element:Cl^{-} atmospheric concentrations. We used the departure from the predicted element to Cl^{-} ratio to determine streamwater enrichment or depletion of an element in drainage waters. Element concentrations greater than the predicted element: Cl^{-} represents a net input from sub-catchment or watershed source pools in the terrestrial ecosystem. Element concentrations below the predicted element:Cl^{-} represent a net loss to the terrestrial ecosystem, or element retention.

5.3.4 Sampling of streamwater pCO_2

We adapted the method of Hope et al. (1995) to sample streamwater in the field using headspace analysis. Stream samples were collected in 1L polypropylene bottles demarcated with a line at 200 mL. The bottle was filled by plunging it into the stream and then filling the bottle completely from the mid-depth flow of streamwater. The excess water was poured off from the full bottle to reach the desired water line (i.e. 200 mL) and the bottle was capped and shaken for 2 min to release dissolved CO_2 in the streamwater into the headspace of the bottle. Three gas samples were taken by inserting a syringe into a septum attached to the top of the sealed sample bottle and extracting 60 mL of gas. This gas sample was then transferred to a glass vial that had a vacuum applied to create a CO_2-free headspace and facilitate the insertion of the sample. The vial was filled by overpressuring the vial from the syringe. Three replicate vials were taken from the bulk sample for analysis in the lab. Stream temperature and pH was measured with an Orion pH/EC probe. Laboratory analyses for pCO_2 were conducted by extracting a 60 mL gas sample from the vial headspace and injecting the sample into a sample loop that was injected into CO_2-free air that carried the sample to a LiCor 7000 infrared gas analyzer. The analyzer returned a peak response that was converted to a CO_2 concentration (μatm) by comparison with a standard curve generated by injecting standards of known CO_2 concentrations through the sample loop. The concentration of CO_2 in the water sample was calculated according the equation:

$$\mu mol\ CO_2\ L^{-1}\ water = \mu L\ CO_2\ L^{-1}\ air \times Kh \times F \quad (1)$$

Where Kh is Henry's constant and F is the fugacity. The fugacity was assumed to be 0.997. The temperature dependence of Henry's law constant was calculated according to:

$$kH = 10\text{^}(-(-108.3865-0.01985076T + 6919.53/T - 669365/T^2 + 40.451*Log(T))) \quad (2)$$

where T is the measured stream temperature. The dissolved free CO_2 present in the original bottle sample was then calculated based on the sum of CO_2 in air and water phases in the sample bottle. We assumed that there was minimal redistribution of other carbonate species and degassing of CO_2 at the low pH values in the study streams.

5.3.5 Streamwater charge mass balance

Streamwater molar charges were calculated from measured stream element concentrations. Total cationic and anionic charge for ions in solution were computed based on the mass balance equation: $\sum (2[Ca^{2+}] + 2[Mg^{2+}] + [Na^{+}] + [K^{+}] + [H^{+}] + [NH_4^{+}]) - \sum (2[SO_4^{2-}] + [NO_3^{-}] + [Cl^{-}] + [HCO_3^{-}])$. We calculated HCO_3^{-} concentrations by two methods. 1) We calculated the charge balance and assumed that the sum of Ca^{2+} and Mg^{2+} charge was equivalent to the $[HCO_3^{-}]$, and 2) we used the assumption in equation 1) and then adjusted for estimated organic acid anionic charge by subtracting the organic acid alkalinity from the $2\{[Ca^{2+}]+[Mg^{2+}]\}$ estimate. We calculated the organic acid charge based on Driscoll et al. (1989) using measured streamwater pH and dissolved organic carbon (DOC). An additional adjustment for Al^{3+} was added to the calculated organic acid charge to neutralize some of the negative charge when Al^{3+} concentration measurements were available. There is some uncertainty in the Al^{3+} adjustment as we did not fractionate the species of Al^{3+} and recognize that there are both variable and uncharged Al^{3+} species present in the solution.

5.4 Results

5.4.1 Stream solution chemistry:rock-water interaction

The streamwater element concentrations reveal patterns of rock-water interactions in the sub-catchments and watersheds that varied according to geomorphic aspects of the watersheds and hydropedologic characteristics of the sub-catchments. The streamwater cation concentrations were highest in the upland sub-catchments and decreased in forested wetland and sloping bog sub-catchments (Table 5.2). The Eaglecrest bog had higher average cation concentrations than the other two sloping bog catchments due to episodically high Ca^{2+} and Mg^{2+} concentrations in the streamwater (Table 5.2). The sub-catchments were arrayed into zones established by Gibbs (1970; 1992) for rivers dominated by weathering, precipitation, or evaporation using the major cation and anion concentration measurements in drainage waters from the sub-catchments (Figure 5.2). The chemical signatures of the mainstem watersheds were distributed according to the

dominant hydrogeologic control elucidated by the ecosystem classification (Nowacki et al., 2001; Figure 5.2a). The recently deglaciated McGinnis watershed had the strongest signature of weathering dominance, while the wetland-dominated Peterson watershed tended toward atmospheric inputs. The Eaglecrest watershed, has a mix of weathering and atmospheric influence due to the presence of both freshly exposed alpine areas and uplifted marine benches (Figure 5.2a). In the sub-catchments, the atmospheric signature dominated the sloping bogs and forested wetlands while the uplands were associated with weathering dominance. The trend from weathering dominated uplands to atmospheric influence in the sloping bog and forested wetland sub-catchments was clearly apparent in the progression toward lower TDS and increased Na^+ in the wetlands compared to the upland sub-catchments (Figure 5.2b). The concentration data indicate that the abundance of Ca^{2+} and Mg^{2+} are the main drivers of the relationship among the watersheds, rather than the abundance of Na^+ (Table 5.2).

5.4.2 Element export and stream indicators of rock-water interaction

The dominant watershed and sub-catchment patterns were quantified through input-output evaluation and an analysis of the atmospheric inputs compared to streamwater outputs. The streamwater Na^+:Cl^- among the sub-catchments was close to the predicted Na^+ concentration from atmospheric input. The streamwater Na^+ concentrations were slightly higher than predicted from the Na^+:Cl^- atmospheric inputs revealing a source of Na^+ in the catchments. The Na^+ concentrations had a trend that decreased slightly from uplands to wetlands, (Figure 5.3a) indicating a possible small Na^+ input from drainage waters derived from uplands across the forested wetlands and sloping bogs. The sub-catchments also had K^+ values above the sea-salt correction line indicating a net source of K^+ from terrestrial source pools (Figure 5.3b). There were two unexplained outliers in the K^+ data that had extremely high values compared to predicted input ratios in the forested wetland sub-catchments, but otherwise there was a fairly consistent grouping of the data, similar to Na^+ concentrations. The sub-catchments have Ca^{2+} concentrations in excess of predicted Ca^{2+}:Cl^- indicating a net source of Ca^{2+} from the terrestrial ecosystem. This trend was especially apparent in the upland sub-catchments (Figure 5.3c). The pattern for Mg^{2+} was similar to Ca^{2+}, but the magnitude of the excess, relative to the sea-salt concentration, was lower than the Ca^{2+} (Figure 5.3d). In both cases, the forested wetlands and sloping bogs had trajectories of Ca^{2+}:Cl^- and Mg^{2+}:Cl^- similar to the Na^+:Cl^- patterns relative to the predicted element:Cl^- ratio (Figure 5.3). The pattern indicates that some of the Ca^{2+} and Mg^{2+}

in these systems may have been derived from flow associated with mineral weathering sources into the wetlands from uplands. The trends for the observed vs. predicted element concentrations are consistent with the Gibbs diagram evaluation where upland sub-catchments have elevated potential weathering products (Ca^{2+} and Mg^{2+}), and the wetland sub-catchment Ca^{2+} and Mg^{2+} values are more closely associated with element concentrations derived from atmospheric inputs and upland source water. The Ca^{2+} and Mg^{2+} concentrations were compared to Si concentrations in the streamwater to evaluate the association with an element that is only available from terrestrial source pools. The Ca^{2+} and Mg^{2+} values in sub-catchment streamwater were significantly related to streamwater Si concentrations (Figure 5.4a, b).

The trends in the streamwater cation concentrations in the mainstem drainages reflected the major hydrogeomorphic influence among the watersheds. Peterson creek was influenced by atmospheric inputs and had the lowest element output among the three watersheds (Figure 5.5). McGinnis had values of Ca^{2+}, Mg^{2+}, and K^{+} in excess of predicted element:Cl^{-} , but most notably in Ca^{2+} and K^{+} (Figure 5.5b, c). Peterson appeared closest to steady-state with atmospheric inputs. This pattern was especially strong at higher concentrations of Cl^{-} in the streamwater (Figure 5.5a, d). McGinnis Creek and Eaglecrest Creek had the greatest departure from atmospheric inputs and had high rates of terrestrial inputs of Ca^{2+} and Mg^{2+} to streamwater (Figure 5.5c, d). McGinnis Creek had a high output of K^{+} relative to the other two watersheds (Figure 5.5b).

5.4.3 Charge balance of the streamwaters across the hydropedologic gradient

The relationships between the estimated HCO_3^{-} alkalinity estimation and pH reveal the influence of organic acids on the total streamwater alkalinity (Figure 5.6). The alkalinity estimate was modeled with a sigmoidal curve similar to David and Vance (1991). The significant fit of the equation illustrates the strength of the relationship between pH and the distribution of charge in the streamwaters. The HCO_3^{-} alkalinity is reduced and eventually disappears as the streamwater pH decreases and the influence of the estimated organic acid charge indicates an accumulation of net negative charge in the streamwaters. The stream geochemical balance is dominated by positive charges associated with Ca^{2+} and Mg^{2+}, and thus, the estimate of HCO_3^{-} alkalinity is governed by the concentration of these two species at high pH. The relationship with Ca^{2+} and Mg^{2+} alkalinity shifts to organic acid alkalinity at low pH where DOC is abundant.

5.4.4 Dissolved carbon in streamwaters

The calculated bicarbonate values provide a means to compare trends in dissolved carbon speciation in the sub-catchment and mainstem streams. The calculation of HCO_3^- from the sum of Ca^{2+} and Mg^{2+} at high pH was similar to values bicarbonate values calculated by including organic acid alkalinity. However, the addition of organic acid alkalinity in streams with low pH decreased the concentration values to zero or near zero in the low pH streams, which is more consistent with expected charges in the acidic environment. The wide range of values in the uplands (6.8-70.6 mg L^{-1}) was primarily due to the high charge concentrations at the Eaglecrest upland sub-catchment. All of the wetland sub-catchments had estimated concentrations below 6 mg L^{-1}, except for the Eaglecrest sloping bog, which had high Ca^{2+} and Mg^{2+} concentrations and subsequently high bicarbonate charge estimates (Table 5.3). The mainstem streams had estimated bicarbonate concentrations that ranged from 12.0-22.9 mg L^{-1}. The most notable discrepancy in the estimates for bicarbonate between the two methods was in the Peterson watershed where the organic acid calculation was 5 mg L^{-1} higher than the sum of Ca^{2+} and Mg^{2+}.

All streams have excess pCO_2 ($epCO_2$) concentrations in the waters draining the catchments (Table 5.3). The excess pCO_2 measured in the sloping bog and forested wetland catchment drainage waters was 3-6 times higher than atmospheric CO_2 concentrations (Table 5.3). The streamwater pCO_2 was significantly different in the sloping bog sites compared to the forested wetland and upland catchments ($F_{2,20} = 7.2$; $p = 0.005$; Table 5.3). The pattern of pCO_2 and the corresponding streamwater CO_2 concentrations (i.e. mg L^{-1}) is closely associated with the seasonal temperature trend at the sub-catchments where maximum concentrations of pCO_2 in streamwaters coincide with maximum summer temperatures (Figure 5.7). The maximum concentration measured in the sloping bogs (~4 mg L^{-1}) is lower than the average streamwater concentration for DOC (Chapter 4), but does indicate an export pathway for carbon from the catchments. The pCO_2 of the sub-catchment streamwaters is significantly related to streamwater pH ($F_{1,20} = 5.53$; $P = 0.03$). The significant difference in the pH among the sub-catchments was inversely related to the increasing concentration of pCO_2 from upland catchments to sloping bog catchments ($F_{2,20} = 79.6$; $p < 0.001$; Figure 5.7). The pH values increase from 4.4 to 5.8 as the pCO_2 declines consistent with shifts in the carbonate equilibrium of the drainage waters (Drever, 1988). The mean pH in the sloping bog and forested wetland catchments was over 1 pH unit lower than the upland catchments.

5.5 Discussion

5.5.1 Input-output budgets and indicators of weathering

The catchments are well suited for performing element mass-balance through the application of the input-output budget approach. The watersheds meet the two important criteria outlined by and Price et al. (2005) including the presence of crystalline bedrock and no substantial leakage of flow to groundwater sources. The presence of bedrock and till as C horizon material in the catchments represent relatively slowly if not impermeable horizons that form the aquitard for watersheds in the NCTR. The presence of this impermeable surface creates conditions that allow mass balance calculations due to the small loss of elements through regional groundwater flow.

The patterns of element to Cl^- ratios reveal the strength of rock water interaction among the catchments and main stem rivers. These patterns are consistent with the weathering of alumino-silicate parent material and weathering products of secondary clay minerals in the soils. The upland soils are well-developed Spodosols that indicate a highly weathered system. However, there appears to be a good connection with the underlying bedrock in the watersheds that provides a clear signal of alumino-silicate weathering and rock-water interaction supported by the significant relationship of Si with Ca^{2+} and Mg^{2+} (Figure 5.4). The abundance of Ca^{2+} and Mg^{2+} output is consistent with the loss of carbonates in successional sequences with active weathering (Gorham et al., 1979). The K^+ and Na^+ in the drainage water from the sub-catchments are closer to steady-state conditions with atmospheric inputs than Ca^{2+} and Mg^{2+}. The Na^+ concentrations are most tightly coupled to Cl^- in sea-salt aerosols, which are the most likely source of atmospheric inputs to the sub-catchments and watersheds. The lower overall contribution of K^+ relative to Ca^{2+} and Mg^{2+} is consistent with retention of K^+ through biological cycling (Schlesinger, 1997). The available K^+ also provides a constituent for secondary clay mineral formation such as illite that serves as a sink for K^+ in the sub-catchments (Reynolds and Johnson, 1971; Moore and Reynolds, 1989).

The ecosystem of the coastal temperate rainforest is often referred to as 'young' due to the recent glacial coverage and retreat that has provided soil and vegetation communities that have developed over the last 12-14ky. However, the rainfall and acidic nature of the conifers result in intense soil development, weathering and organic matter accumulation. The clear weathering signal detected from the analysis has not been well-documented across several

watershed types in the NCTR. The distinction in the small catchment array demonstrates the disparity in ecosystem maturity and development with regard to weathering interpretations. The NCTR provides a paradox where well-developed soils occur in near surface horizons, but bedrock and till are accessible to drainage waters. The translocation of organic acids through complexation with Al^{3+} and Fe^{3+} creates Spodosols, while carbonic acid acts to weather parent material for export to surface water pathways. The potential for the overall carbon cycling in this system is greatly expanded when both organic and inorganic sources of carbon are considered.

5.5.2 Organic acid alkalinity and stream chemical balance

Soil and streamwater chemical charge balance is heavily influenced by organic acid alkalinity (Urban et al., 1989; White and Shannon, 1997). Organic acid charge can be a significant portion of the overall balance in streams due to the activity of organic acid functional groups. The presence of peatlands and organic acids creates conditions where excess negative charge accumulates in stream waters (David and Vance, 1991). Streamwaters draining peatlands have measured negative charge of up to -58 $\mu Eq\ L^{-1}$ as charge densities on organic acids can average 48 μmol_c per mmol DOC (White and Shannon, 1997). Therefore, our approach of correcting for the presence of organic acids at low pH values provides a more accurate estimate of the stream alkalinity.

There are also several functional attributes to consider given the presence of the organic acids in the solution. The dissolved organic matter (DOM) may play an important role in mediating electron transfers in the streamwater. Exported soil organic matter represents a large storage pool of energy as reduced compounds in the terrestrial environment. This pool of energy is transformed by electron transfers that process material through bacterial metabolism (Lovley, et al., 1996; Scott et al., 1998). The implications of high amounts of reactive organic matter in the stream systems of the NCTR are clear, but as yet unresolved. Observations of DOM quality quantified the labile nature of the DOM from wetlands (Fellman et al., 2008; 2009). An injection experiment demonstrated that this labile DOM could be respired by stream organisms over short distances (Fellman et al., 2009). It is clear from these studies that the reactive DOM interacts with the stream soon after exiting the soil matrix and is a component of the pCO_2 dynamics of the streams.

5.5.3 Inorganic carbon export from catchments

The concentration of dissolved CO_2 in the catchments is consistent with the cycle of biological production and respiration increases in soil pore waters (Hope et al., 2001). The concentration pattern is consistent across the hydrologic gradient confirming a biological source for at least part of the dissolved CO_2 concentrations in the streams. It is clear that a portion of the CO_2 produced in the soils is not captured as soil respiration but leaves the soil matrix via soil flow paths. The response of soil respiration in the wetlands must be viewed with the potential for alternative export pathways.

The concentrations of dissolved CO_2 are not extremely high compared with other tributary and mainstem rivers systems (Dawson et al., 2009; Table 5.4). The concentrations have a large range, with increased dissolved CO_2 fluxes during stormwater flow from the soils. Catchment carbon budgets in the NCTR must be able to capture this vector of carbon flux to adequately capture the total carbon budget. The concentration discharge relationships were not strong enough to develop highly predictive curves for dissolved inorganic carbon (DIC) flux in the uplands, and only weak relationships in the wetlands. The range of concentrations can be compared to other similar systems (Table 5.4) to estimate the potential flux for use in total carbon flux models. The concentration of the dissolved CO_2 can be compared to the dissolved carbon concentration in the hydropedologic types to estimate the total annual dissolved carbon flux. The concentration is approximately 20% of the DOC flux, leading to a potential for export of < 5 g m^{-2} y^{-1} from the bogs and forested wetlands, and considerably less from the uplands. These values are fairly small compared to the DOC fluxes (Chapter 4), and most of this dissolved CO_2 will likely evade from the tributaries or mainstem rivers leading to an eventual atmospheric loss from the system.

The presence of low alkalinity waters with high pCO_2 concentrations has implications for the near shore coastal ecosystems. The flow of acidic river drainage waters has been implicated in the disruption of the geochemical balance of coastal waters that influence the production of shell in marine organisms (Salisbury et al., 2008). The need to quantify the outputs from catchments source pools that combine in river output to the estuaries of northern temperate systems is especially important due to the abundance of low alkalinity waters. There is evidence for reduced DOC flux due to warming in some interior Alaskan rivers (Striegl et al., 2005). The trajectory of material outputs in the NCTR is not well understood, but an increase in DOC alkalinity at the

expense of bicarbonate alkalinity would have a negative impact on the shell-bearing organisms of the coastal zone.

5.6 Conclusions

The rock-water interaction reveals an inorganic perspective for carbon cycle dynamics in NCTR watersheds. Input-output budgets provide a means to estimate HCO_3^- and organic acid alkalinity to explain potential implications of charge balance in streamwaters draining individual sub-catchments distinguished by hydropedological types. There is a clear weathering signature in upland sub-catchments among the well-drained hydropedologic types. Biological production in the poorly-drained wetland hydropedologic types promotes dissolved CO_2 production and export while diminishes the weathering signature and export of inorganic carbon as HCO_3^-. The potential for inorganic carbon export and movement of free CO_2 coupled with the demonstrated abundance of DOC (Chapter 4) makes the soil-stream linkage compelling for overall ecosystem carbon cycling. The present active cycling and potential sequestration of carbon highlights the role that NCTR forests and peatlands play in the regional carbon budget. Future climate shifts are also a distinct highlight as small changes in temperature and precipitation may exert a great deal of leverage in the balance of the inorganic and organic components of the carbon cycle. There are a multitude of possible outcomes for the fate of carbon due to the feedback of the climate drivers on the organic and inorganic terrestrial carbon cycles, which are closely coupled with aquatic systems.

5.7 References

Alaback PB (1996) Biodiversity patterns in relation to climate: the coastal temperate rainforests of North America Chapter 7 In: Lawford RG, Alaback PB, Fuentes E (eds) High-latitude rainforests and associated ecosystems of the west coast of the Americas. Ecological Studies 116, Springer-Verlag, New York pp, 105-133.

Berner EK, Berner RA (1987). The global water cycle, geochemistry, and environment. Prentice-Hall, Inc. Englewood Cliffs NJ.

Birkeland P (1984) Soils and Geomorphology. Oxford, UK, Oxford University Press.

Bormann FH, Likens G (1979) Pattern and process in a forested ecosystem. Springer-Verlag, New York.

Caldeira K (2006) Forests, climate, and silicate rock weathering. J Geochem Explor 88:419-422.

Chandler RF (1942) The time required for podzol formation as evidenced by the Mendenhall glacial deposits near Juneau, Alaska. Soil Sci Soc Am Proc 7:454-459.

Chapin FS, Walker LR, Fastie CL, Sharman LC (1994) Mechanisms of primary succession following deglaciation at Glacier Bay, Alaska. Ecol Monogr 64:149-175.

Christophersen N, Seip H, Wright RF (1982) A model for streamwater chemistry at Birkenes, Norway. Water Resour Res 18: 977-996.

Crews T, Kitayama K, Fownes JH, Riley RH, Herbert DA, Mueller-Dombois D, Vitousek PM (1995) Changes in soil phosphorus fractions and ecosystem dynamics across a long chronosequence in Hawaii. Ecology 76(5):1407-1424.

D'Amore DV, Fellman JB, Edwards RT, Hood E (2010) Controls on dissolved organic matter concentrations in soils and streams from a forested wetland and sloping bog in southeast Alaska. Ecohydrol 3:249-261.

David MB, Vance GF (1991) Chemical character and origin of organic acids in streams and seepage lakes of central Maine. Biogeochemistry 12:17-41.

Dawson JC, Billett MF, Hope D, Palmer SM, Deacon CM (2004) Sources and sinks of aquatic carbon in a peatland stream continuum. Biogeochemistry 70:71-92.

Dawson JC, Soulsby CS, Hrachowitz M, Speed M,. Tetzlaff D (2009) Seasonality of $epCO_2$ at different scales along an integrated river continuum within the Dee Basin, NE Scotland. Hydrol Proc 23:2929-2942.

Drever JI (1988) The geochemistry of natural waters. Prentice Hall, Englewood Cliffs NJ.

Driscoll CT, Fuller RD, Schecher WD (1989) The role of organic acids in the acidification of surface waters in the eastern U.S. Water Air Soil Poll 43:21-40.

Feller MC (1977) Nutrient movement through western hemlock-western redcedar ecosystems in southwestern British Columbia. Ecology 58:1269-1283.

Fellman JB, D'Amore DV, Hood E, Boone RD (2008) Fluorescence characteristics and biodegradability of dissolved organic matter in forest and wetland soils from coastal temperate watersheds in southeast Alaska. Biogeochemistry 88:169-184.

Fellman JB, Hood E, D'Amore DV, Edwards RT (2009) Seasonal changes in the chemical quality and biodegradability of dissolved organic matter exported from soils to streams in coastal temperate rainforest watersheds. Biogeochemistry 95:277-293.

Gibbs RJ (1970) Mechanisms controlling world water chemistry. Science 170: 1088-1090.

Gibbs RJ (1992) A reply to comments of Eilers et al. Limnol Oceanogr 37:1338-1339.

Gorham E, Vitousek PM, Reiners WA (1979) The regulation of chemical budgets over the course of terrestrial ecosystem succession. Ann Rev Ecol Syst 10:53-84.

Hedin LO, Hetherington E (1995) Atmospheric and geologic constraints on the biogeochemistry of North and South American temperate rain forests. In: Lawford RG, Fuentes E, Mooney H (eds) High latitude forest and riverine systems of the west coasts of the Americas: environment, ecology and human use. Springer-Verlag, New York.

Hedin LO, Armesto JJ, Johnson AH (1995) Patterns of nutrient loss from unpolluted old-growth temperate forests: evaluation of biogeochemical theory. Ecology 76(2): 493-509.

Holmes RM, McClelland JW, Raymond PA, Frazer BB, Peterson BJ, Stieglitz M (2008) Lability of DOC transported by Alaskan rivers to the Arctic Ocean. Geophys Res Lett 35, L03402. Doi:10.1029/2007GL032837.

Hope D, Dawson JC, Cresser MS, Billett MF (1995) A method for measuring free CO_2 in upland streamwater using headspace analysis. J Hydrol 166:1-14.

Hope D, Palmer S, Billett MF, Dawson JJC (2001) Carbon dioxide and methane evasion from a temperate peatland stream. Limnol Oceanogr 46:847-857.

Ice GG, Stednick JD (2004) Forest watershed research in the United States. Forest History Today (Spring/Fall): 16-26.

Johnson NM (1971) Mineral equilibria in ecosystem geochemistry. Ecology 52(3): 529-531.

Johnson NM, Likens GE, Bormann FH, Pierce RS (1968) Rate of chemical weathering of silicate minerals in New Hampshire. Geochimica et Cosmochimica Acta 32(5): 531-538.

Kennedy MJ, Chadwick OA, Vitousek PM, Derry LA, Hendricks DM (1998) Replacement of weathering with atmospheric sources of base cations during ecosystem development, Hawaiian Islands. Geology 26:1015-1018.

Lin H (2003) Hydropedology: bridging disciplines, scales, and data. Vadose Zone J 2:1-11.

Lin H, Bouma J, Pachepsky Y, Western A, Thompson J, van Genuchten R, Vogel H J, Lilly A (2006) Hydropedology: synergistic integration of pedology and hydrology. Water Resour Res 42:W05301, doi:10.1029/2005WR004085.

Lovley DR, Coates JD, Blunt-Harris EL, Phillips EJP, Woodward JC (1996) Humic substances as electron acceptors for microbial respiration. Nature 382:445-448.

Lugo AE, Swanson FJ, Gonzalez OR, Adams MB, Palik B, Thill RE, Brockway DG, Kern C, Woodsmith R, Musselman R (2006) Long-term research at the USDA Forest Service's experimental forests and ranges. Bioscience 56(1): 39-48.

Miller RD (1973) Gastineau Channel formation: a composite glaciomarine deposit near Juneau, Alaska. Geological Survey Bulletin, contributions to stratigraphy, USGS Report #1394, Washington DC.

Moore DM, RCJ Reynolds (1989) X-ray Diffraction and the Identification and Analysis of Clay Minerals. Oxford University Press New York.

NADP (2011) National Atmospheric Deposition Program/National Trends Network website: http://nadp.sws.uiuc.edu/.

NACP (2011) North American Carbon program, website: http://www.nacarbon.org/nacp/.

Nowacki G, Shephard M, Krosse P, Pawuk W, Fisher G, Baichtal J, Brew D, Kissinger E, Brock T (2001) Ecological subsections of southeast Alaska and neighboring areas of Canada. Anchorage AK, US Department of Agriculture Forest Service, Alaska Region 10, R10-TP-75, 306 pp.

Palmer SM, Hope D, Billett MF, Dawson JC, Bryant CL (2001) Sources of organic and inorganic carbon in a headwater stream: evidence from carbon isotope studies. Biogeochemistry 52:321-338.

Perakis SS, Hedin LO (2007) State factor relationships of dissolved organic carbon and nitrogen losses from unpolluted temperate forest watersheds. J Geophys Res 112:G02010, doi:10.1029/2006JG000276, 2007.

Price JR, Velbel MA, Patino LC (2005) Rates and time scales of clay-mineral formation by weathering in saprolitic regoliths of the southern Appalachians from geochemical mass balance. GSA Bull 117(5/6):783-794.

Reuss J, Johnson D (1986) Acid deposition and the acidification of soils and waters. Ecological Investigations 59, Springer-Verlag New York.

Reynolds RC, NM Johnson (1971) Chemical weathering in the temperate glacial environment of the Northern Cascade Mountains. Geochimica et Cosmochimica Acta 36:537-554.

Richter DD, Hofmockel M, Callaham MA, Powlson DS, Smith P (2007) Long-term soil experiments: keys to managing Earth's rapidly changing ecosystems. Soil Sci Soc Am J 71:266-279.

Salisbury J, Green M, Hunt C, Campbell J (2008) Coastal acidification by rivers: a threat to shellfish? EOS 89:513-514.

Schlesinger W (1997) Biogeochemistry: an analysis of global change. Academic Press, San Diego.

Scott DT, McKnight NM, Blunt-Harris EL, Kolesar SE, Lovley DR (1998) Quinone moieties act as electron acceptors in the reduction of humic substances by humics-reducing microorganisms. Environ Sci Technol 32:2984-2989.

Soil Survey Division Staff (1999) Soil Taxonomy: a basic system of soil classification for making and interpreting soil surveys. US Government printing office, AH-436, Second edn., Washington, DC.

Soil Survey Staff (2010) Keys to soil taxonomy 11th edn. USDA Natural Resources Conservation Service, Washington.

Stevenson F (1982) Humus Chemistry. John Wiley and Sons, New York.

Stednick JD (1981) Precipitation and streamwater chemistry in an undisturbed watershed in southeast Alaska. USDA Forest Service, Pacific Northwest Research Station, Research Paper PNW-291.

Streigl RG, Aiken GR, Dornblaser MM, Raymond PA, Wickland KP (2005) A decrease in discharge-normalized DOC export by the Yukon River during summer through autumn. Geophys Res Lett 32, L21413, doi:10.1029/2005GL024413.

Sugai SF, Burrell DC (1984) Transport of dissolved organic carbon, nutrients, and trace metals from the Wilson and Blossom rivers to Smeaton Bay, southeast Alaska. Can J Fish Aq Sci 41:180-190.

Sverdrup H, Warfvinge P (1988) Weathering of primary silicate minerals in the natural soil environment in relation to a chemical weathering model. Water, Air, Soil Pollut 38: 387-408.

Sverdrup H, Warfvinge P (1991) Calculating critical loads of acid deposition with PROFILE: a steady-state soil chemistry model. Water Air Soil Pollut 63: 119-143.

Urban NR, Bayley SE, Eisenreich SJ (1989) Export of dissolved organic carbon and acidity from peatlands. Water Resour Res. 25:1619-1628.

Vitousek PM, Farrington H (1997). Nutrient limitation and soil development: experimental test of a biogeochemical theory. Biogeochemistry 37: 63-75.

Vitousek PM, Rieners W (1975) Ecosystem succession and nutrient retention: a hypothesis. Bioscience 25: 376-381.

Wallin M, Buffam I, Oquist M, Laudon H, Bishop K (2010) Temporal and spatial variability of dissolved inorganic carbon in a boreal stream network: concentrations and downstream fluxes. J Geophys Res 115:G02014, doi:10.1029/2009jg001100.

White JR, Shannon RD (1997) Modeling organic solutes in peatland soils using acid analogs. Soil Sci Soc Am J 61:1257-1263.

Worrall F, Swank WT, Burt T (2005) Fluxes of inorganic carbon from two forested catchments in the Appalachian Mountains. Hydrol Proc 19:3021-3035.

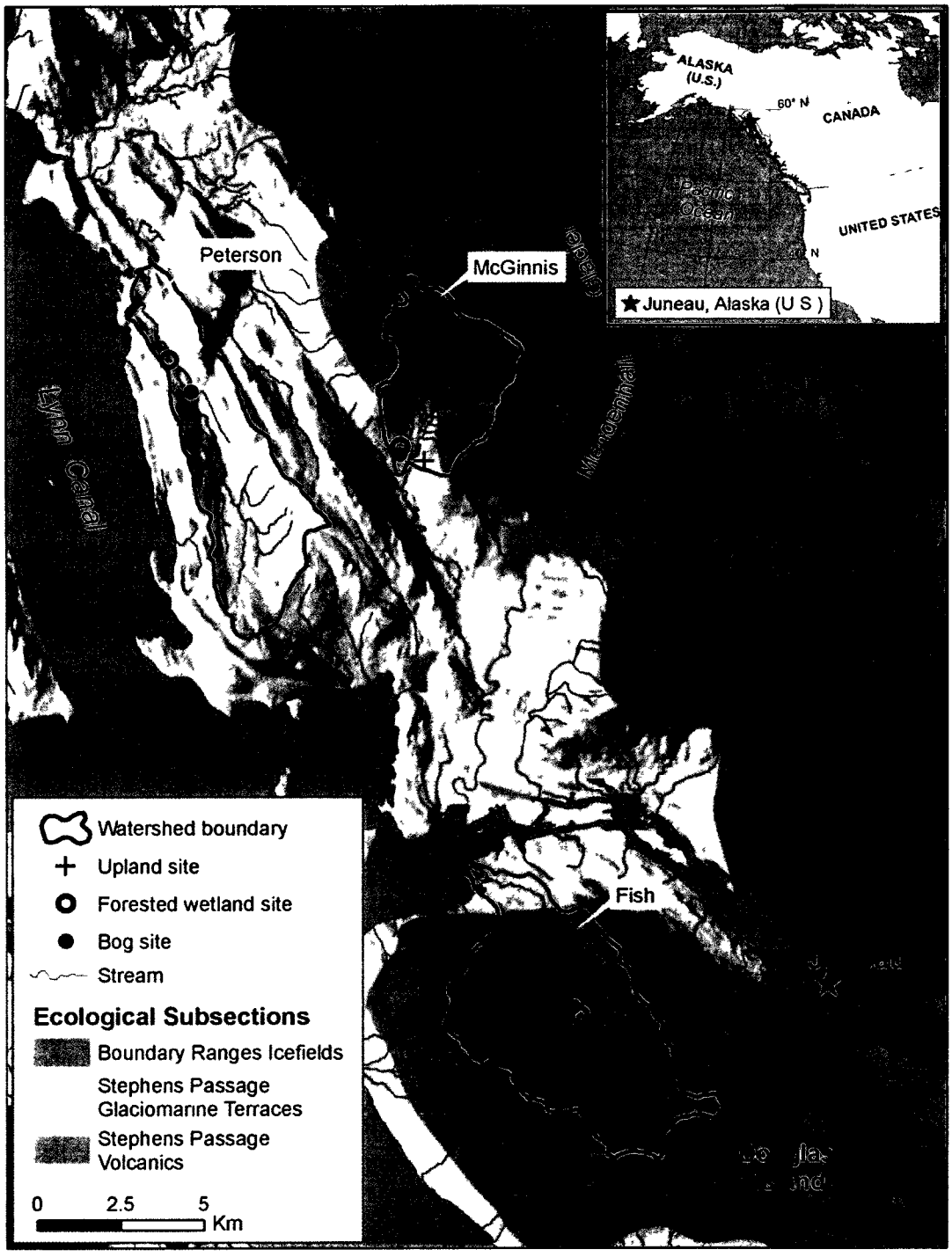

Fig. 5.1. The ecological subsections and watersheds of the Juneau, AK area. The approximate locations for measurements of stream chemistry in sloping bogs, forested wetlands, and uplands are identified within the watersheds (Peterson, McGinnis, and Eaglecrest/Fish).

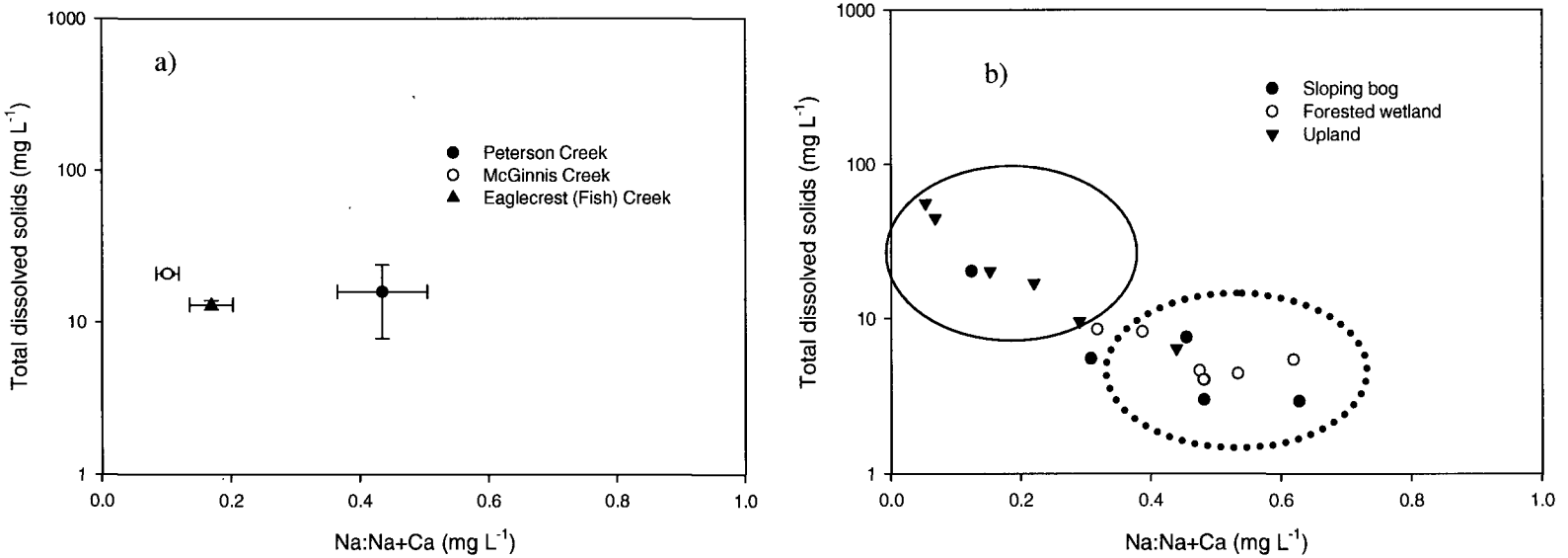

Figure 5.2. Sources of dissolved ions in mainstem streams and in outlet streams of hydropedologic types. Identification of source for dissolved ions in stream water from the mainstem streams (a) are derived from weathering sources in McGinnis and Eaglecrest/Fish creeks and tend toward atmospheric sources in Peterson Creek. Within hydropedologic types (b), dissolved ions in waters draining uplands are derived from weathering (solid ellipse) while those in sloping bogs and forested wetlands are derived primarily from atmospheric sources (dotted ellipse). Diagram modified from Gibbs (1970).

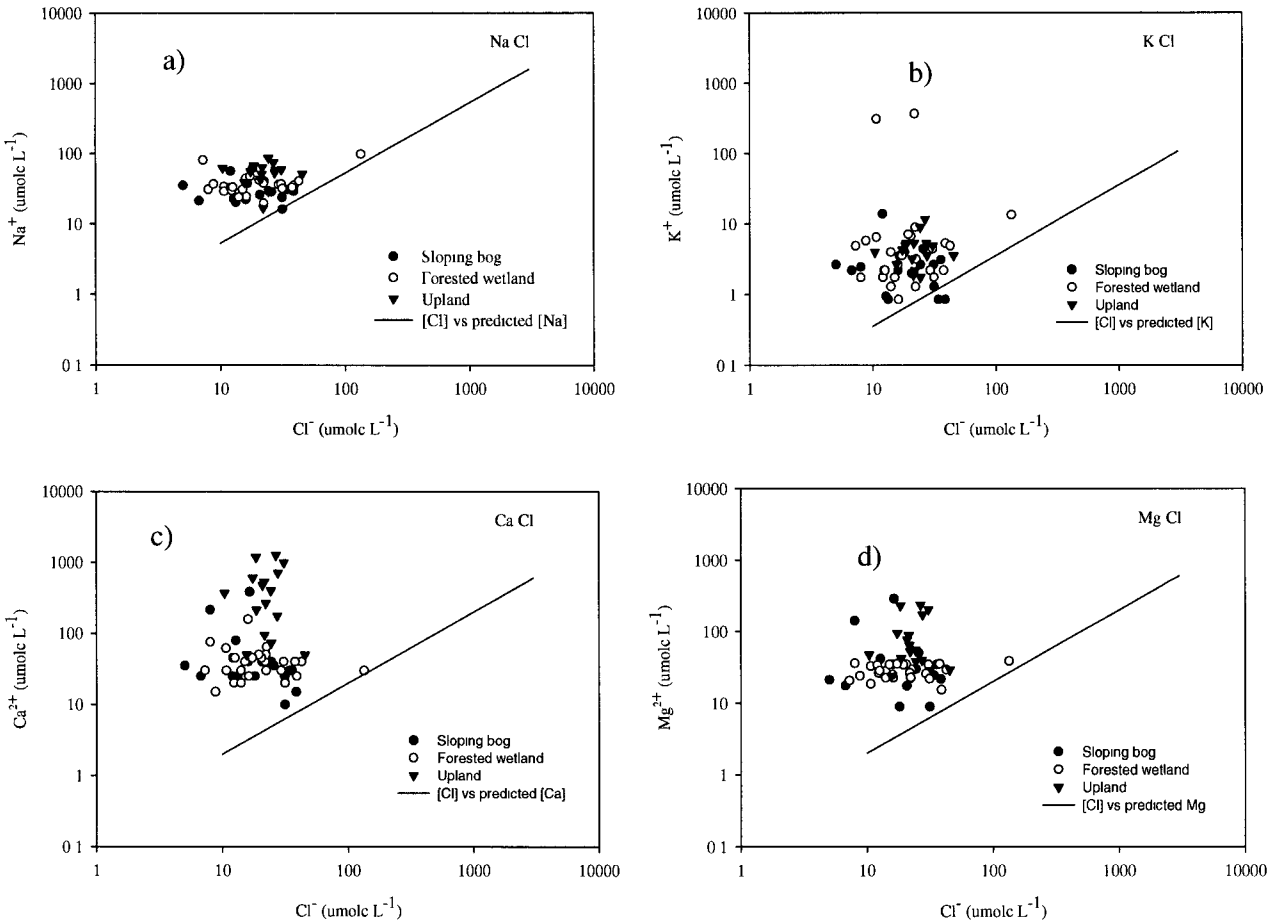

Figure 5.3. Streamwater concentrations of dissolved cation to chloride ratios in hydropedologic types. Solid lines represent predicted cation:Cl^- in atmospheric inputs for comparison to measured outputs in drainage waters.

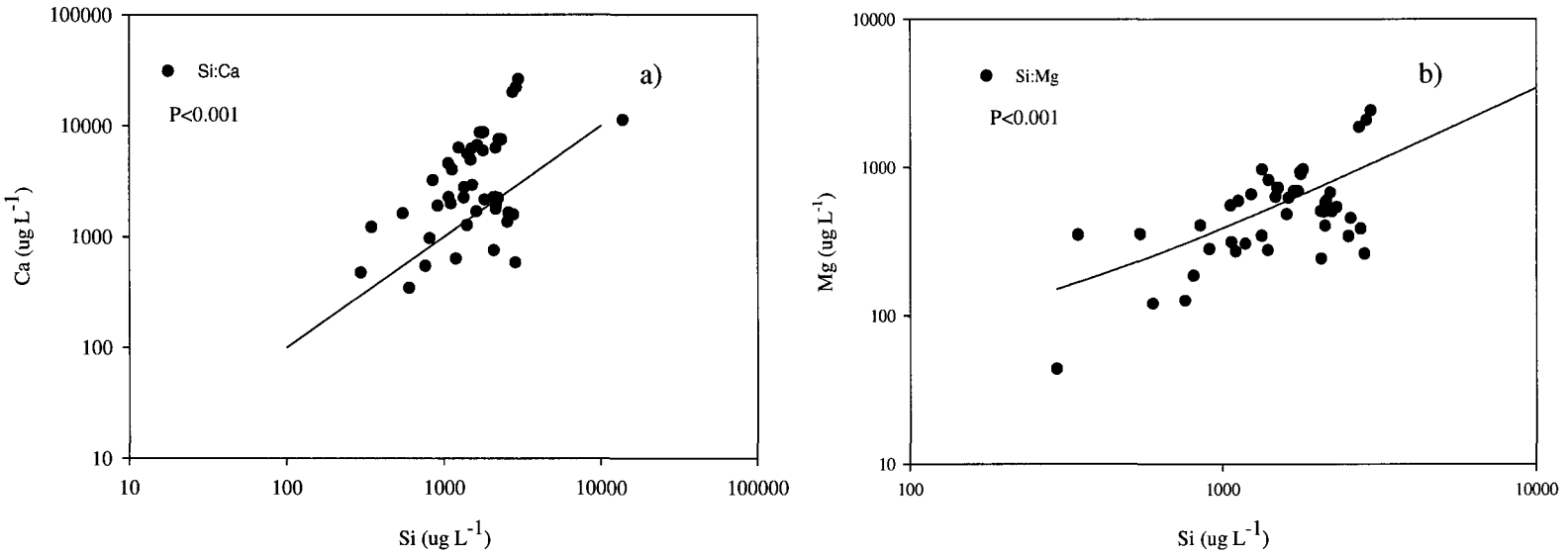

Figure 5.4. Correlations of silica to magnesium and calcium in stream waters of hydropedologic types. Ca^{2+} (a) and Mg^{2+} (b) are significantly related to Si in stream waters draining hydropedologic types.

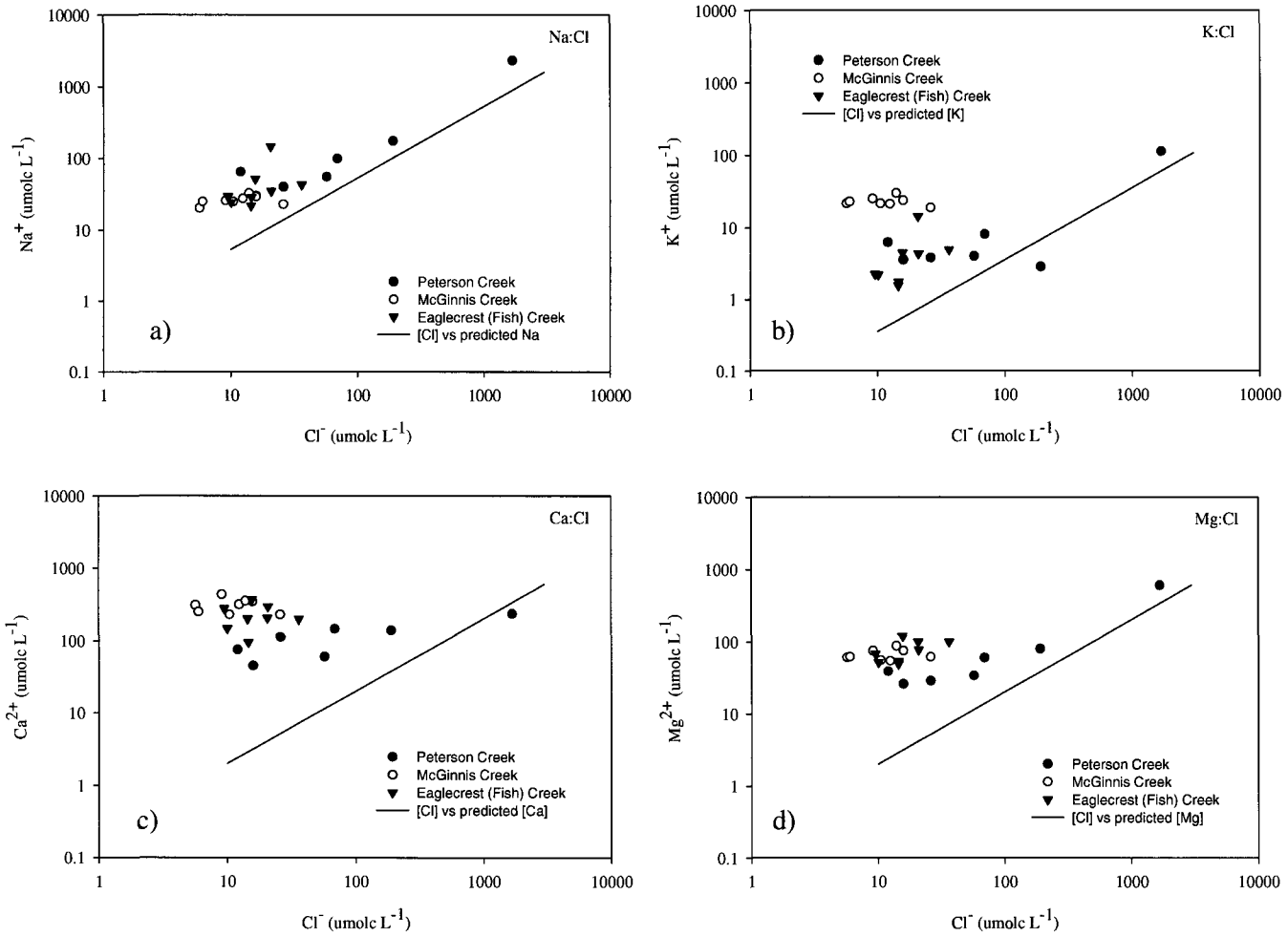

Figure 5.5. Streamwater concentrations of dissolved cation to chloride ratios in mainstem streams. Solid lines represent predicted cation:Cl^- in atmospheric inputs for comparison to measured outputs in drainage waters.

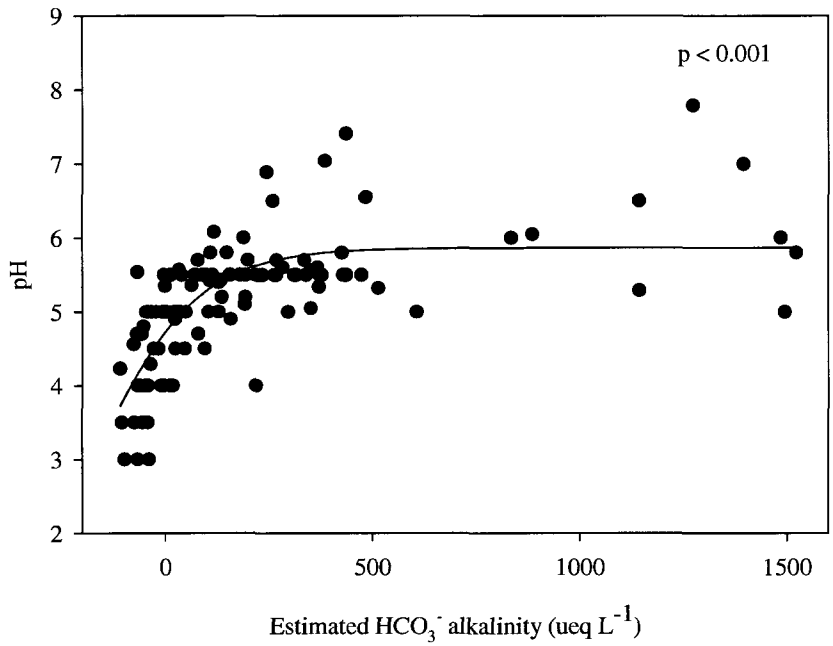

Figure 5.6. Relationship between estimated alkalinity and pH in drainage waters of hydropedologic types and mainstem streams.

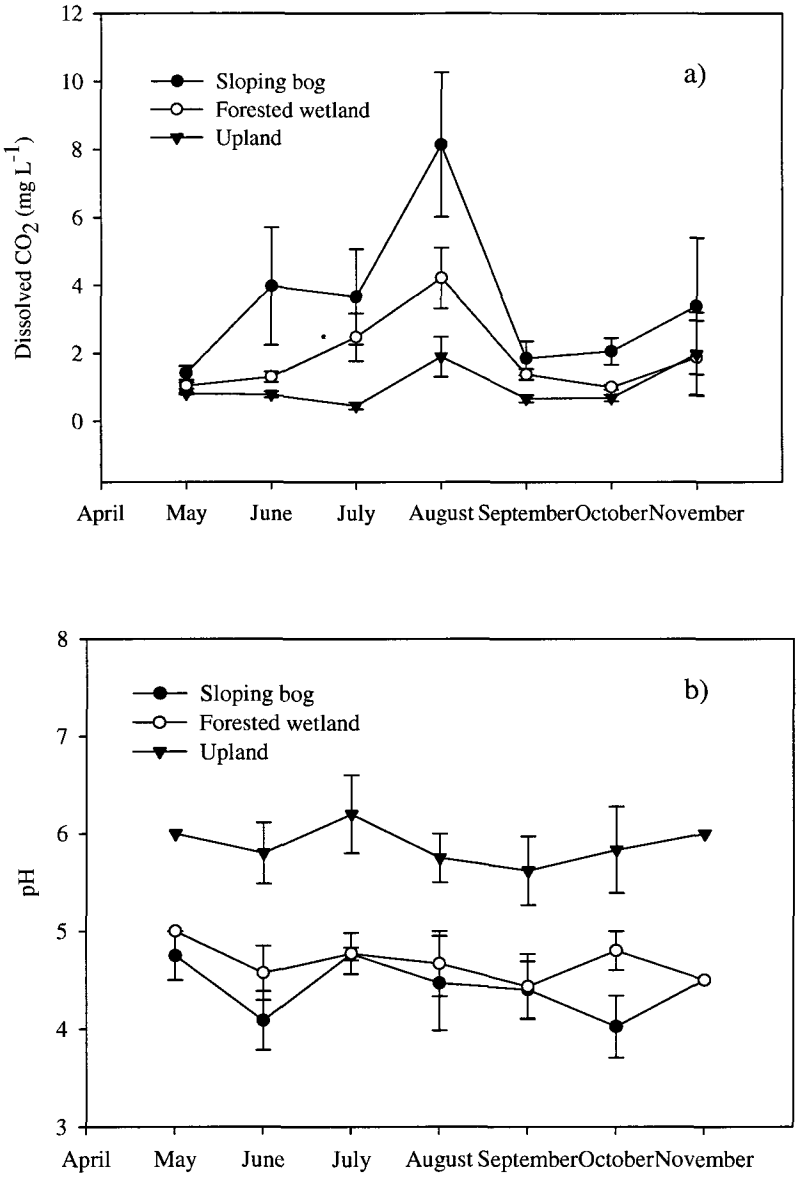

Figure 5.7. Dissolved CO_2 and pH in drainage waters of hydropedologic types. Concentrations of a) dissolved CO_2, and b) pH in waters draining hydropedologic types (mean ± SE n = 3 per hydropedologic type).

Table 5.1. Soil classification and attributes of hydropedologic types within three watersheds in the Juneau, AK area.

Site	Ecosystem type	Ecosystem code	Taxonomic classification[†]	Elevation (m)	Slope (%)	Drainage class[§]
Eaglecrest	Bog	EBT	Dysic Typic Cryohemist	89	3	Very poorly-drained
	Forested Wetland	EFW	Coarse-loamy, mixed, dysic Terric Cryohemists	248	5	Poorly-drained
	Upland	EUT	Coarse-loamy, mixed, superactive, Typic Haplocryod	392	15	Well-drained
McGinnis	Bog	MBT	Dysic Typic Cryohemist	133	3	Very poorly-drained
	Forested Wetland	MFW	Loamy, mixed, Terric Cryohemists	128	19	Somewhat poorly-drained
	Upland	MUT	Coarse-loamy, mixed, superactive, Typic Humicryod	202	20	Well-drained
Peterson	Bog	PBT	Dysic Typic Cryohemist	112	2	Very poorly-drained
	Forested Wetland	PFW	Coarse-loamy, mixed, Histic Cryaquept	20	5	Poorly-drained
	Upland	PUT	Loamy, mixed, superactive, Lithic Haplocryod	163	20	Well-drained

[†]Taxonomic classification represents the field and laboratory interpretation for the site, not the official series.

[§]Drainage class was determined by field observations and soil map unit assignment from soil resource inventory.

Table 5.2. Elevation, area, and concentrations of major cations and anions in mainstem streams and waters draining hydropedologic types‡.

Watershed	Elevation	Area	Ca^{2+}	Mg^{2+}	Na^{+}	K^{+}	SO_4^{-}	Cl^{-}	Si
Hydropedologic type	(m)	(ha)	----------mg L^{-1}----------						
Peterson	931	2646	2.65	0.68	2.40	0.32	0.37	0.64	1.39
Sloping bog	112	1	1.37	0.33	0.61	0.05	0.14	0.58	0.76
Forested wetland	20	5	1.52	0.40	0.96	0.15	0.12	0.58	2.61
Upland	163	20	2.16	0.54	0.88	0.09	0.3	0.82	2.10
McGinnis	2236	1690	7.08	0.77	0.56	0.88	1.46	0.32	1.50
Sloping bog	133	0.5	0.59	0.12	0.55	0.06	0.07	0.33	0.57
Forested wetland	128	5	0.63	0.21	0.57	0.06	0.06	0.37	1.88
Upland	202	1	7.13	0.48	1.28	0.17	0.78	0.59	2.22
Eaglecrest	1563	3539	5.19	0.78	0.66	0.10	0.38	0.53	1.43
Sloping bog	89	1	4.7	1.98	0.66	0.08	0.06	0.56	5.57
Forested wetland	248	10	1.89	0.58	0.88	0.07	0.17	0.51	1.97
Upland	392	8	22.85	2.13	1.28	0.20	1.01	0.71	2.85

Table 5.3. Measurements of pH and dissolved organic and inorganic carbon in mainstem streams and waters draining hydropedologic types.

Watershed Hydropedologic type	pH	DOC ($mg\ L^{-1}$)	$epCO_2$ (ppmv)	CO_2 (μatm)	$HCO_3^{-\dagger}$	$HCO_3^{-\ddagger}$
					---------($mg\ L^{-1}$)--------	
Peterson	5.2	12.79	0.59	984 (106)	14.60	19.14
Sloping bog	4.7	14.77	2.73	1037 (140)	4.78	0.92
Forested wetland	5.1	15.75	2.60	988 (186)	4.77	3.68
Upland	5.2	7.68	0.05	726 (246)	7.65	6.77
McGinnis	5.7	15.70	2.50	951 (103)	22.87	19.97
Sloping bog	3.9	18.51	3.64	1383 (219)	2.49	0.00
Forested wetland	4.0	24.49	2.41	916 (143)	4.56	1.32
Upland	5.9	3.69	1.39	529 (79)	22.05	22.19
Eaglecrest	5.7	9.79	2.54	966 (182)	18.36	17.67
Sloping bog	4.6	13.89	5.38	2046 (609)	15.94	12.00
Forested wetland	5.0	10.11	1.76	668 (56)	5.36	4.25
Upland	6.4	5.11	1.25	474 (30)	70.57	69.09
All sloping bog	4.4 (3.0-5.0)		3.7 (1.8-8.2)	1533 (324-7112)		
All forested wetland	4.6 (3.0-5.6)		2.3 (1.4-4.5)	882 (393-3215)		
All upland	5.8 (5.0-7.0)		1.5 (1.0-2.3)	539 (305-1557)		

$^{\dagger}HCO_3^-$ calculated by $2\{[Ca^{2+}]+[Mg^{2+}]\}$.

‡ HCO_3^- calculated by $(\sum Cations-\sum Anions)-(RCOO^-) = HCO_3^-$.

Table 5.4. Inorganic carbon concentrations and flux in streamwater studies of several ecosystem types and within hydropedologic types in southeast Alaska.

	Streamwater concentration ($mg\ L^{-1}$)		Annual area specific CO_2 and HCO_3^- export ($g\ C\ m^{-2}\ y^{-1}$)		Annual area specific DIC export ($g\ C\ m^{-2}\ y^{-1}$)	
Ecosystem type	CO_2	HCO_3^-	CO_2	HCO_3^-		Reference
Temperate forest	0.9-4.9	---	1.68-14.17	4.25	9.7-23.9	Worrall et al., 2005
NCTR	---	---	---	16.0	---	Stednick, 1981
Boreal streams	---	---	---	---	0.1-1.5	Wallin et al., 2010
Temperate heath	0.3-1.8	0.8-1.1	---	---	---	Palmer et al., 2001
Temperate heath	0.3-1.9	0.1-2.4	0.3-1.0	0.03-0.7	---	Dawson et al., 2004
Temperate heath	0.7-3.1	---	---	---	---	Hope et al., 1995
Temperate forest	---	5.6-10.0	---	---	---	Feller, 1977
Sloping bog	0.3-19.5	0-15.9	---	---	---	This study
Forested wetland	0.5-8.6	1.3-5.4	---	---	---	This study
Upland	0.2-4.4	6.8-70.6	---	---	---	This study

Chapter 6: A carbon budget model for coastal temperate rainforest watersheds in southeast Alaska[1]

6.1 Abstract

Forests play an important role in the global balance of carbon flows between atmospheric and terrestrial pools. Many carbon budgets focus almost exclusively on the transfer of carbon between terrestrial vegetation and the atmosphere through gaseous fluxes. However, the large pool of soluble carbon that is transferred via running waters from the terrestrial ecosystem also plays a role in some humid regions, and total carbon balances in wet forests may not be accurate unless the flow of dissolved carbon is accounted for in budget models. The North American perhumid coastal temperate rainforest (NCTR) of the North Pacific coast has some of the largest terrestrial carbon stocks in the world. We devised a conceptual model for estimating the carbon balance of three NCTR watersheds based on the assignment of carbon fluxes from terrestrial ecosystem types to distinct hydrologic classes. Estimates of carbon flux were based on two approaches that used soil drainage class as a predictive variable. The first approach assigned three classes based on divisions of soil map units into low, moderate, or high flux categories for soil respiration or dissolved organic carbon. The second approach applied functions derived from a continuous regression function across the hydrologic gradient to capture classes in between the major intermediate divisions. We combined estimates for net primary production from remote sensing with empirical equations for terrestrial carbon fluxes developed from extensive sampling within the research watersheds. Estimates of net ecosystem production (NEP) ranged from 1.13 to 2.69 Mg C ha^{-1} with an average of 2.04 ± 0.81 Mg C ha^{-1} among the watersheds. There was very little difference (<1%) between the two estimation methods, and the carbon accumulation rates indicate that the watersheds were net carbon accumulators between 2000 and 2009. The uncertainty in NEP estimates for any specific year is high, but the conceptual approach of using soil drainage as a controlling factor for carbon flux worked well as a method for the NCTR watersheds. The carbon budget is the first estimate of landscape NEP for the NCTR and highlights the importance of dissolved carbon fluxes in the total ecosystem carbon budget.

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[1] D'Amore DV, Edwards RT, Rupp TS, Biles FE, Nay SM (to be submitted to Ecosystems) A carbon budget model for coastal temperate rainforest watersheds in southeast Alaska.

6.2 Introduction

Forest carbon storage is an important component of the carbon balance where the global stock of terrestrial carbon is estimated to be 2.3 Pg (Denman et al., 2007). The flux of carbon through this stock is complicated, yet essential to understanding the role of forests in global carbon cycles. Temperate forests contribute to the terrestrial carbon store at a rate of approximately 2.1 Pg y^{-1} (Houghton, 2003). It is believed that old-growth forests are important carbon stocks, while young-growth forests are important sites for carbon sequestration. However, the strength of the sink in old-growth forests is believed to be as important or more important than managed landscapes (Harmon et al., 1990) and the concept that old forests retain their sink strength longer than originally thought has been proposed (Luyssaert et al., 2008). The carbon dense region of the Northeastern Pacific Coastal margin has become a focus for carbon cycling research due to the potential for maintaining large accumulations of carbon in old-growth and accumulating carbon in young growth forests.

The estimation of landscape carbon budgets requires both thorough empirical data that quantify the various components of the carbon flux (Harmon et al., 2004) and the application of spatially explicit estimates of carbon flux (Schimel, 1995). Verification of estimates on a regional basis is an important component of the quality control for global carbon flux estimation. Scaling plot measurements up across varying terrain and ecosystem types is a challenge in plot-based work in regional carbon budget programs. Stratification of output budgets is often done by climatic zone (Campbell and Law, 2005), stand composition, or harvest history (Davis et al., 2009). This allows for a landscape stratification among stand types over time. Any approach requires detailed carbon accounting that includes estimates of net primary productivity (NPP), soil respiration, dissolved fluxes, and other export vectors of carbon for estimation across scales (Pendall et al., 2004; Reichstein and Beer, 2008).

Forest carbon inventories take two approaches to measure carbon flow, using either plot inventory or remote sensing. Landscape carbon inventories can utilize dispersed forest inventory information for inputs of forest biomass (Jenkins et al., 2001; Van Tuyl et al., 2005; Hudiburg et al., 2009). Input budgets are constructed through estimation of components of biomass and application of allometric relationships to determine the fate of biomass in various pools. Outputs are measured in time series to develop predictive equations for carbon fluxes from vegetation and soils (Law et al., 2003). Forest carbon outputs rely on spatially explicit measurements of fluxes of soil respiration, and in some cases other components such as dissolved organic carbon (DOC),

volatile organic carbon (VOC), and herbivory (Chapin et al., 2005). Carbon accumulation in forest stands is highly dependent on forest age (Pregitzer and Euskirchen, 2004). Therefore, landscape approaches to forest carbon balance must account for age-related aspects of forest growth combined with ecosystem and climate variables. The North American perhumid coastal temperate rainforest (NCTR) is subject mostly to small-scale disturbance that creates a mosaic of stand ages that is difficult to capture with field survey techniques in a rigorous way to account for the total landscape variability. The variability in ecosystem types that range from densely forested stands to open bogs is also a dramatic small-scale feature of the landscape (Neiland, 1971).

Net ecosystem production needs to account for the various fate of carbon accumulated in an ecosystem, including DOC (Randerson et al., 2002; Lovett et al., 2006). The dominant flux in northern temperate watersheds is through soil respiration (Ryan and Law, 2005), but in extremely wet ecosystems such as rainforests, outputs such as dissolved forms of carbon (i.e. DOC and dissolved inorganic carbon, DIC) must be considered to complete the net ecosystem carbon balance (Chapin et al., 2005).

Carbon accounting is difficult due to the many levels of uncertainty in measurements. One key factor is the representation of variability on the landscape. Average estimates obtained from point-specific data must be generalized across the overall landscape. There is a need in many regions for capturing the variability in terrain, which is especially acute in forests. The forests of the NCTR are highly variable due to the microtopography and shifts in soil moisture.

Soil moisture tends to limit productivity in forests of some regions, but excess soil moisture leads to soil saturation and decreases in growth rates of aboveground biomass. Therefore, capturing the hydrologic gradient provides a great deal of interpretive power across the landscape.

The NCTR provides a challenge in deriving carbon budgets due to the variation in vegetation productivity and carbon flux across the hydropedologic gradient. Variation in stand age also leads to varying and complicated combinations of inputs and outputs. The production and export of carbon as DOC must also be considered given the high rates of flux from wetlands into surface water pathways (Chapter 4). We applied estimates of the major inputs and outputs in the carbon budget across the diverse landscape of three NCTR watersheds to determine the range of carbon storage. The three watershed types represent broad geomorphic groupings and provide an opportunity to examine how empirically derived carbon flux estimates combined with an

estimate of net primary productivity can be used to predict carbon budgets across a heterogeneous landscape.

6.3 Materials and methods

6.3.1 Experimental approach

We used two methods to calculate the carbon balance in three experimental watersheds of the NCTR (Figure 6.1; Table 6.1). The first approach assigned values to discrete landscape units based on empirically derived estimates among three ecosystem types: sloping bogs, forested wetlands, and uplands. These three ecosystem types contain soils that are the main functional units along the hydrologic gradient in the NCTR. We use the term "hydropedologic type" to define the combination of soil and hydrology into functional units (Chapter 2). Soil map units within three research watersheds were identified from Tongass National Forest soil resource inventory (USDA Forest Service, 1997; Figure 6.1). Each of the soil map units also had a corresponding drainage class based on soil morphologic attributes on a scale of 1-5 from poorly-drained to well-drained. Each soil map unit was assigned a value for soil respiration and DOC flux according to fluxes calculated from the three drainage class end-members (Chapters 3 and 4). The soil drainage classes between 1 and 2 were assigned to sloping bog, 2 and 3 to forested wetland, and 4 and 5 to upland. The second approach constructed predictive equations for soil respiration and DOC flux across the drainage classes to provide a continuous response and error associated with the prediction. This approach allowed a comparison of the variability in estimates by binning, and in addition, provided an estimate of the uncertainty in the carbon flux estimates for soil respiration and DOC within each drainage class.

6.3.2 Soil respiration rates

Soil respiration measurements were taken on collars that were permanently placed within each hydropedologic type (sloping bogs, forested wetlands, and uplands) in the three experimental watersheds (Chapter 3). Soil respiration measurements were done using a dynamic-closed chamber procedure without drawdown (Nay et al., 1994). The collars were designed to receive a portable closed-chamber field respirometer that used a LiCor 820 infra-red gas analyzer (LiCor, Inc., Lincoln, NE) for a total of 21 collars per catchment type. Soil respiration measurements were taken every 2 to 4 weeks during the spring, summer, and fall over a three-year period from 2007-2009. All soil respiration measurements were accompanied by a soil

temperature measurement (10 cm depth) at each collar during each measurement. Annual fluxes were estimated by combining temperature dependent models with measured and estimated soil temperatures. The contribution of root respiration to total soil respiration was assumed to be 50% based on the overall mean reported in a review of the contribution of root respiration to soil respiration (Hanson et al., 2000).

6.3.3 Dissolved organic carbon export

Streamwater DOC flux was calculated from continuous discharge measurements combined with intermittent streamwater samples of DOC concentration (Chapter 4). Concentration-discharge relationships and estimation of DOC flux were calculated using the load estimator program LoadEst (Runkel et al., 2004). Data input and output was facilitated by use of the Loadrunner program (Raymond and Saiers, 2010). Estimates of load (i.e. DOC flux) were based on adjusted maximum likelihood estimation (see Runkel et al., 2004 for details). The calculation of load is based on a non-linear regression calculated from known DOC concentration and associated measurements of stream discharge as dependent and independent variables. Estimates of annual flux were calculated by interpolation from the concentration discharge relationship that computed flux for days that did not have streamwater DOC concentration data. Daily estimates were summed for each watershed and area-weighted fluxes were derived by distributing the total annual flux over the watershed area of the catchment.

6.3.4 Estimate of particulate organic carbon

Particulate organic carbon (POC) was not measured in the study. Therefore, we used published values of DOC:POC and estimated the contribution of POC based on the watershed export of DOC. The values had a broad range, with varying relationships across ecosystem types. However, there were not apparent trends that could be readily assigned to the landscape units by function. We calculated the mean ratio based on the published values of Hope et al. (1994).

6.3.5 Estimate of dissolved inorganic carbon

The annual DIC flux was in the watersheds was estimated through a predictive relationship between concentrations and flux of HCO_3^- and pCO_2 in streamwaters from published values (Worrall et al., 2005). The relationship established to predict flux from concentration was then applied to the calculated streamwater HCO_3^- and pCO_2 concentrations (Chapter 5) to derive an annual DIC flux in the experimental watersheds. The annual flux estimate was then applied to

the entire watershed through assignment by soil map unit (smu) and similar to DOC and soil respiration. The estimates for the drainage class (dc) model are the same as the sum model as there was no reliable method to build predictive relationships through the dc model. Therefore, the DIC estimates do not vary between the two models.

6.3.6 Calculation of net primary productivity

The MODIS satellite project (MOD17) provided a gridded dataset for NPP estimates at a 1 km resolution. The MOD17 algorithm is a continuous satellite-driven dataset monitoring global vegetation productivity (Zhao et al., 2005). We used the 2000-2006 annual average NPP dataset calculated by the Numerical Terradynamic Simulation Group (NTSG) to estimate watershed NPP. The NTSG produced this estimate by filling cloud contaminated pixels, refining the spatial resolution of the meteorological input data, and recalibrating the biome look-up table used in the MOD17 algorithm (Zhao et al., 2005). The algorithm is based on the assumption that NPP is linearly related to the amount of absorbed photosynthetically active radiation (PAR) during the growing season. Grid cells covering non-vegetated areas (permanent wetland, barren or sparsely vegetated, urban, and water body) are classified as "no data" and are not included in NPP estimates.

6.3.7 Conceptual carbon budget model

6.3.7.1 Basic model (all values in g C m^{-2} y^{-1})

$$NEP = NPP - [Rh + DOC + POC + DIC]$$

NEP: Net ecosystem productivity

NPP : Net primary productivity

Rh: Heterotrophic soil respiration

DOC: Dissolved organic carbon

POC: Particulate organic carbon

DIC: Dissolved inorganic carbon

6.3.7.2 Derivation of NPP

NPP = MODIS estimate by watershed

6.3.7.3 Derivation of heterotrophic respiration

$Rh = Rt - Ra$

Rt: Total soil respiration efflux

Ra: Autotrophic respiration, assumed to be half of Rt on an annual basis

6.3.7.4 Dissolved organic carbon flux

$DOC = Qs \times [DOC]$

Qs = specific discharge, $L\ s^{-1}$

DOC concentration = $mg\ L^{-1}$

6.3.8 Estimates for values of model components from equations based on landscape responses across drainage gradient

The drainage gradient is represented by soil map unit drainage classes that range from 1-5 (see methods). Drainage classes were assigned by average water table depth in 10 cm increments. Water table depth 0-10 = dc 1, very poorly-drained; 10-20 cm = dc 2, poorly-drained; 20-30 cm = dc 3, somewhat poorly-drained; 30-40 cm = dc 4, moderately well-drained; 40-50 cm = dc 5, well-drained.

6.3.8.1 Basic equation for NEP derived for each landscape soil map unit

$NEP_{smu} = NPP_{smu} - [Rh_{smu} + DOC_{smu}]$

NEP_{smu}: Net ecosystem productivity in a soil map unit

NPP_{smu}: Net primary productivity in a soil map unit

Rh_{smu}: Heterotrophic soil respiration export

DOC_{smu}: Dissolved organic carbon

6.3.8.2 Derivation of heterotrophic respiration by drainage class

$Rh = Rt - Ra$

Rt: Total soil respiration efflux

Ra: Autotrophic respiration, estimated to be half of Rt on an annual basis

$Rt_{smu} = \beta 0 * \exp(-\beta 1 * Rt_{dc}) + \varepsilon$

6.3.8.3 Dissolved organic carbon flux by drainage class

$$DOC = Qs \times [DOC]$$

Qs = specific discharge, L s^{-1}

DOC concentration = mg L^{-1}

$$DOC_{smu} = \beta 0 * (DOC_{dc})^{\beta 1} + \varepsilon$$

6.3.8.4 Watershed net ecosystem productivity:

$$WS_{nep\ (smu,dc)} = NPP - \sum_{k=5}^{n} \text{WS Flux out}_{smu,dc}$$

With,

$$NEP_{smu,dc} = NPP - \sum_{k=5}^{n} (Rh_{smu,dc} + DOC_{smu,dc})$$

Where,

$WS_{nep\ (smu,\ dc)}$ = Total watershed net ecosystem productivity by soil map unit or drainage class

NPP = Estimated watershed net primary productivity

$Rh_{smu,\ dc}$ = Estimated heterotrophic respiration from each soil map unit or drainage class

$DOC_{smu,\ dc}$ = Estimated dissolved organic carbon flux from each soil map unit or drainage class

6.4 Results

6.4.1 Estimates of NPP from remote sensing

The MOD17 improved NPP estimates varied among the watersheds coincident with the range of ecosystem types located within the watershed (Table 6.2). The highest mean NPP rate was measured in the Peterson watershed. This estimate was influenced somewhat by the amount of coverage in forest and the highest estimated minimum NPP among the watersheds. The minimum NPP estimates in both McGinnis and Eaglecrest watersheds were most likely due to extensive alpine areas in both watersheds. The maximum NPP was estimated in the Eaglecrest watershed where mature valley bottom forests dominate the lower reaches of Fish Creek, the main drainage in the watershed. The maximum NPP in Peterson and McGinnis watersheds were not much lower. The mean values were all below 400 g m^{-2} y^{-1}, at the low range of values for forests of the Pacific Northwest where values are often >500 g C m^{-2} y^{-1} (Hudiburg et al., 2009).

6.4.2 Soil respiration

The soil respiration rates were closely associated with drainage gradient. Soil respiration estimates ranged from 71-235 g C m^{-2} y^{-1}. The rates within ecosystem type grouped very closely with very low variability in estimates of temperature dependent soil respiration fluxes (Chapter 3). The close association of the respiration rates was driven by the close control of soil and vegetation community selection. The association between drainage class assigned by water table depth had a strong trend across the drainage gradient, but also some large variability (Figure 6.2). The three sloping bogs grouped closely in drainage class 1, with soil respiration rates ranging from 71-85 g C m^{-2} y^{-1}. The Peterson forested wetland site in drainage class 1.5 had an annual soil respiration rate more closely associated with the forested wetland sites in drainage class 2.5 than the sloping bogs in drainage class 1. The Peterson forested wetland had a mean soil respiration rate of 149 g C m^{-2} y^{-1}, which was similar to the other two forested wetlands assigned to drainage class 2.5 where soil respiration rates were slightly higher (171 and 179 g C m^{-2} y^{-1}). The soil respiration rate in the Peterson forested wetland site is likely to be related to the presence of overstory trees that contributed to autotrophic respiration and fine root turnover.

The respiration rates among the uplands in drainage classes 4.5 and 5 varied between 194 and 235 g C m^{-2} y^{-1}. The McGinnis upland in drainage class 4.5 had a respiration rate similar to the forested wetlands in drainage class 2.5. The large gap between drainage classes 2.5 and 4.5, in soil respiration response leads to some uncertainty in the landscape estimation. There is the potential for an ecosystem in drainage classes 2-5 to have a total soil respiration flux between 180 and 220 g C m^{-2} y^{-1}. Similarly, the uncertainty in the soil respiration rate between drainage classes 1 and 2 had a range between 71 and 149 g C m^{-2} y^{-1}

6.4.3 Dissolved organic carbon

The DOC fluxes in drainage classes 1, 1.5, and 2.5 were all greater than 20 g C m^{-2} y^{-1}, while drainage classes 4.5 and 5 were less than 15 g C m^{-2} y^{-1} (Figure 6.3). The large discrepancy in the DOC estimates among the drainage classes gives a great deal of leverage to low drainage classes in the predictive model (Figure 6.3) and the cumulative DOC flux from wetlands on the landscape. The flux of DOC decreased across the hydropedologic gradient from wet to dry consistent with the carbon stock present in the soils. The averages in hydropedologic type decreased from 33 and 31 g C m^{-2} y^{-1} in sloping bogs and forested wetlands to approximately 8 g C m^{-2} y^{-1} in uplands (Chapter 4, Figure 4.3). The influence of the leverage of wetland class smu

on total DOC flux was apparent in the estimates in Peterson watershed where the average watershed flux (29 g C m^{-2} y^{-1}) was closer to the average value for wetlands than uplands (Table 6.3). The estimate in the Peterson watershed for DOCsmu was also the most similar to the DOCdc estimate among the watersheds (Table 6.3).

6.4.4 Net ecosystem productivity

The estimates of NEP provide evidence that the watersheds accumulated carbon in terrestrial stocks. The average NEP among the watersheds was 2.4 ± 0.8 Mg C ha^{-1}. Estimates based on NEPsmu vs. NEPdc did not vary greatly. Estimates using either method resulted in similar values for NEP in the watersheds with differences approximately 1% higher for NEPsmu in the Eaglecrest and Peterson watersheds, and 1% lower for NEPsmu than NEPdc in the McGinnis watershed (Table 6.4). McGinnis watershed had the lowest NEP values for both NEPsmu and NEPdc while Peterson watershed had the highest values (Table 6.4).

The estimates of additional outputs that included POC and DIC decreased the overall carbon storage in the watersheds (Table 6.4). The decrease in the NEP estimate was approximately 6-18% across the watersheds. The greatest reduction occurred in Eaglecrest due to the large estimate of HCO_3^- export from the uplands. The decreases in overall NEP did not greatly diminish the value and the estimates still showed a considerable amount of net storage in the watersheds.

6.4.5 Comparison of NPP and NEP

The ratio of NEP to NPP is approximately 0.50-0.75 in the forests. The McGinnis watershed is lowest with 0.50 followed by the Eaglecrest (0.70) and Peterson (0.75) watersheds. These estimates represent a high accumulation rate by either the NEPsmu or NEPdc method of estimation. The rate of the McGinnis watershed is similar to the predicted NEP:NPP relationship, while both Eaglecrest and Peterson are 15 and 20% higher than the ratio of 0.56 (Figure 6.4; Pregitzer and Euskirchen, 2004).

6.5 Discussion

6.5.1 Assessment of uncertainty in the model

6.5.1.1 NPP

The NPP estimate has several challenges that make accurate estimates problematic in the NCTR. The largest obstacle to accurate measurements of NPP during the growing season is the common occurrence of cloud cover that obscures the ground and consequently leads to a lack of information in many of the pixels. The interpolation value in the estimate (i.e. QC) represents a metric that is related to the percentage of 8-day leaf area index (LAI) values that were filled during the growing season due to unfavorable atmospheric conditions (see Zhao et al., 2005). These QC values for the NCTR are relatively high compared to other regions and lead to less reliable estimates using the remote sensing MODIS17 approach. The mosaic of terrestrial vegetation types creates a dilemma for adequate sampling in any one ecosystem type. For example, plot-based estimates can lead to bias upward in the estimation of NPP from biomass increment (Jenkins et al., 2001) due to the often non-random selection of plots and trees for sampling. Despite the drawbacks of the MODIS17 approach, the remotely sensed information provides one of the best available estimates of values for NPP across an entire watershed in the NCTR. Therefore, the broad estimates across the watershed provide some assurance that the value is representative of the sum of ecosystem components.

6.5.1.2 Soil respiration and DOC models

The estimates of soil respiration and DOC at both extremes of soil saturation have a much higher degree of certainty than the intermediate levels of soil saturation. Soil respiration estimates vary between 70-150 g C m^{-2} y^{-1} in drainage class 1. This variability is mainly due to the presence of one site with high water table and high respiration rates. The result is a slightly higher mean value than the apparent response of the similar sites (i.e. sloping bogs). The two sites at the extreme end of the drainage sequence in drainage class 5 are well within the confidence intervals for the estimate. Both the values at drainage class 2 and 3 are either on the border of the confidence interval or outside the confidence interval. The lack of a value for drainage class 4 leads to a great deal of uncertainty in this range. One factor driving the variability of soil drainage class is the fewer site-specific measurements of water table compared to the wide distribution of

sampling for soil respiration. The wider distribution of sampling for soil respiration appears to lead to more consistent site means.

The similar pattern of uncertainty is present in the predictive relationship for DOC where extreme measurements have a much smaller range of variability than the intermediate values. In addition, the lack of measurement data for drainage class 4 makes it impossible to use the prediction for site-specific estimates of either soil respiration or DOC. The consistency in the DOC flux measurements is achieved with repeated measurements in time, but not space, similar to the water table measurements. However, the integrating nature of streamflow is apparent in the similarity of landscape units across the hydrologic gradient.

6.5.1.3 Potential for estimates of watershed carbon sequestration

The mean estimates for NEP indicate that the watersheds are accreting carbon in biomass and soils. The magnitude of the estimate must be considered carefully, but the consistent large net accretion in the watersheds across the two models indicates that the measured forests are actively sequestering carbon. The estimates are similar to young-growth stands in the Pacific Northwest where mature forest stands (80-200 y) had NEP rates of 2.86 Mg C ha^{-1} y^{-1} and old forests (>200 y) had rates of 1.65 Mg C ha^{-1} y^{-1} (Turner et al., 2000). The gradient of NEP rates among the experimental watersheds is consistent with the trend of lower accumulation rates in younger forests on the continental sites at McGinnis and Peterson watersheds and the mature sites in the Eaglecrest watershed. These patterns reflect the relationship of forest age to annual carbon storage where initial losses change to rapid carbon accretion after canopy closure (Law et al., 2003; Gough et al., 2008). The carbon accretion curve model by Leighty et al. (2006) estimates carbon accretion rates of 2.0-5.0 Mg C ha^{-1} y^{-1} in forests up to 250 years old based on forest inventory sampling.

The ratio of NPP to NEP is higher compared to assessments of other forested ecosystems (Figure 6.4). This may indicate an overestimate of NEP in the NCTR watersheds. There is a good indication that the trajectory of NEP is positive or a net sink for carbon in the forested ecosystem given the magnitude of the potential overestimate. The indication that the forests are sinks for carbon is consistent with large-scale estimates that demonstrate northern latitude forests as carbon sinks (Houghton, 2003). The abundant store of woody debris and deep pools of soil carbon do indicate that the forest has been a net sink for carbon during the Holocene.

6.6 Conclusions

The conceptual drainage class model is a promising approach for estimating carbon budgets in the NCTR. The partitioning of the landscape and assignment of component fluxes provided reasonable estimates for NEP using both discrete and continuous model approaches. The watershed estimates indicate that the terrestrial ecosystem is accreting carbon consistent with values for early (100-300 y) conifer forest development. The approach using the soil saturation gradient provides a method to address the continuous nature of terrestrial ecosystem heterogeneity to estimate carbon balance. At large scales, this approach can provide preliminary estimates of carbon balance while continued measurements can improve the accuracy of the predictions.

6.7 References

Campbell JL, Law BE (2005) Forest soil respiration across three climatically-distinct chronosequences. Biogeochemistry 73:109-125.

Chapin FS III, Woodwell GM, Randerson JT, Lovett GM, Rastetter EB, Baldocchi DD, Clark DA, Harmon ME, Schimel DS, Valentini R, Wirth C, Aber JD, Cole JJ, Giblin A, Goulden ML, Harden JW, Heimann M, Howarth RW, Matson PA, McGuire AD, Melillo JM, Mooney HA, Neff JC, Houghton RA, Pace ML, Ryan MG, Running SW, Sala OE, Schlesinger WH, Schulze E-D (2005) Reconciling carbon-cycle concepts, terminology, and methodology. Ecosystems 9:1041-1050.

Davis SC, Hessl AE, Scott CJ, Adams MB, Thomas RB (2009) Forest carbon sequestration changes in response to timber harvest. For Ecol Manage 258:2101-2109.

Denman KL, Brasseur G, Chidthaisong A, Ciais P, Cox PM, Dickinson RE, Hauglaustaine D, Heinze, Holland C, Jacob E, Lohmann D, Ramachandran U, DaSilva S, Dia PL, Wofsy SC, Zhang, X (2007) Couplings between changes in the climate system and biogeochemistry. In: Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, Miller HL (eds) Climate change 2007: the physical science basis. Contribution of working group I to the fourth assessment report of the intergovernmental panel on climate change. Cambridge University Press, Cambridge, UK/New York, pp. 499-587.

Gough CM, Vogel CS, Schmid HP, Curtis PS (2008) Controls on annual forest carbon storage: lessons from the past and predictions for the future. Bioscience 58:609-622.

Hanson PJ, Edwards NT, Garten CT, Andrews JA (2000) Separating root and soil microbial contributions to soil respiration: a review of methods and observations. Biogeochemistry 48:115-146.

Harmon ME, Ferrell WK, Franklin JF (1990) Effects on carbon storage of conversion of old-growth to young forests. Science 247:699-702.

Harmon ME, Bible K, Ryan MG, Shaw DC, Chen H, Klopatek J, Li X (2004) Production, respiration, and overall carbon balance in an old-growth *Pseudotsuga-Tsuga* forest ecosystem. Ecosystems 7:498-512.

Hope D, Billett MF, Cresser MS (1994) A review of the export of carbon in river water: fluxes and processes. Environ Pollut 84:301-324.

Houghton RA (2003) Why are estimates of the terrestrial carbon balance so different? Glob Change Biol 9:500-509.

Hudiburg T, Law B, Turner DP, Campbell J, Donato D, Duane M (2009) Carbon dynamics of Oregon and Northern California forests and potential land-based carbon storage. Ecol Appl 19:163-180.

Jenkins JC, Birdsey RA, Pan Y (2001) Biomass and NPP estimation for the mid-Atlantic region (USA) using plot-level forest inventory data. Ecol Appl 11:1174-1193.

Law BE, Sun OJ, Campbell J, Van Tuyl S, Thornton PE (2003) Changes in carbon storage and fluxes in a chronosequence of ponderosa pine. Glob Change Biol 9:510-524.

Leighty WW, Hamburg SP, Caouette J (2006) Effects of management on carbon sequestration in forest biomass in southeast Alaska. Ecosystems 9:1051-1065.

Lovett GM, Cole JJ, Pace ML (2006) Is net ecosystem production equal to ecosystem carbon accumulation? Ecosystems 9:152-155.

Luyssaert S, Schulze ED, Borner A, Knohl A, Hessenmoller D, Law BE, Ciais P, Grace J (2008) Old-growth forests as global carbon sinks. Nature 455:213-215.

Nay SM, Mattson KG, Bormann BT (1994) Biases of chamber methods for measuring soil CO_2 efflux demonstrated with a laboratory apparatus. Ecology 75:2460–2463.

Neiland BJ (1971) The forest-bog complex in Southeast Alaska. Vegetatio 22:1-63.

NTSG (2011) Numerical Terradynamic Simulation Group, www.ntsg.umt.edu/project/mod17.

Pendall E, Bridgham S, Hanson PJ, Hungate B, Kicklighter DW, Johnson DW, Law BE, Luo Y, Megonigal JP, Olsrud M, Ryan MG, Wan S (2004) Below-ground process responses to elevated CO_2 and temperature: a discussion of observations, measurement methods, and models. New Phytol 162:311-322.

Pregitzer KS, Euskirchen ES (2004) Carbon cycling and storage in world forests: biome patterns related to forest age. Glob Change Biol 10:1-26. Doi:10.1111/j.1365-2486.00866.x.

Randerson JT, Chapin FS, Harden JW, Neff JC, Harmon ME (2002) Net ecosystem production: a comprehensive measure of net carbon accumulation by ecosystems. Ecol Appl 12:937-947.

Raymond PA, Saiers JE (2010) Event controlled DOC export from forested watersheds. Biogeochemistry doi10.1007/s10533-010-9416-7.

Reichstein MC, Beer C (2008) Soil respiration across scales: the importance of a model-data integration framework for data interpretation. J Plant Nutr Soil Sci 171:344-354.

Ryan MG, Law BE (2005) Interpreting, measuring, and modeling soil respiration. Biogeochemistry 73:3-27.

Runkel RL, Crawford CG, Cohn TA (2004) Load estimator (LOADEST): a FORTRAN program for estimating constituent loads in streams and rivers: U.S. Geological Survey Techniques and Methods Book 4, Chapter A5, 69 pp.

Schimel DS (1995) Terrestrial ecosystems and the carbon cycle. Glob Change Biol 1:77-91.

Turner DP, Cohen WB, Kennedy RE (2000) Alternative spatial resolutions and estimation of carbon flux over a managed landscape in Western Oregon. Landscape Ecol 15:441-452.

USDA Forest Service (1997) Tongass Land Management Plan. United States Forest Service, Alaska Region.

Van Tuyl S, Law BE, Turner DP, Gitelman AI (2005) Variability in net primary production and carbon storage in biomass across Oregon forests-an assessment integrating data from forest inventories, intensive sites, and remote sensing. Forest Ecol Manage 209:273-291.

Worrall F, Swank WT, Burt T (2005) Fluxes of inorganic carbon from two forested catchments in the Appalachian Mountains. Hydrol Proc 19:3021-3035.

Zhao M, Heinsch FA, Nemani RR, Running SW (2005) Improvements of the MODIS terrestrial gross and net primary production global data set. Remote Sens Environ 95:164-176.

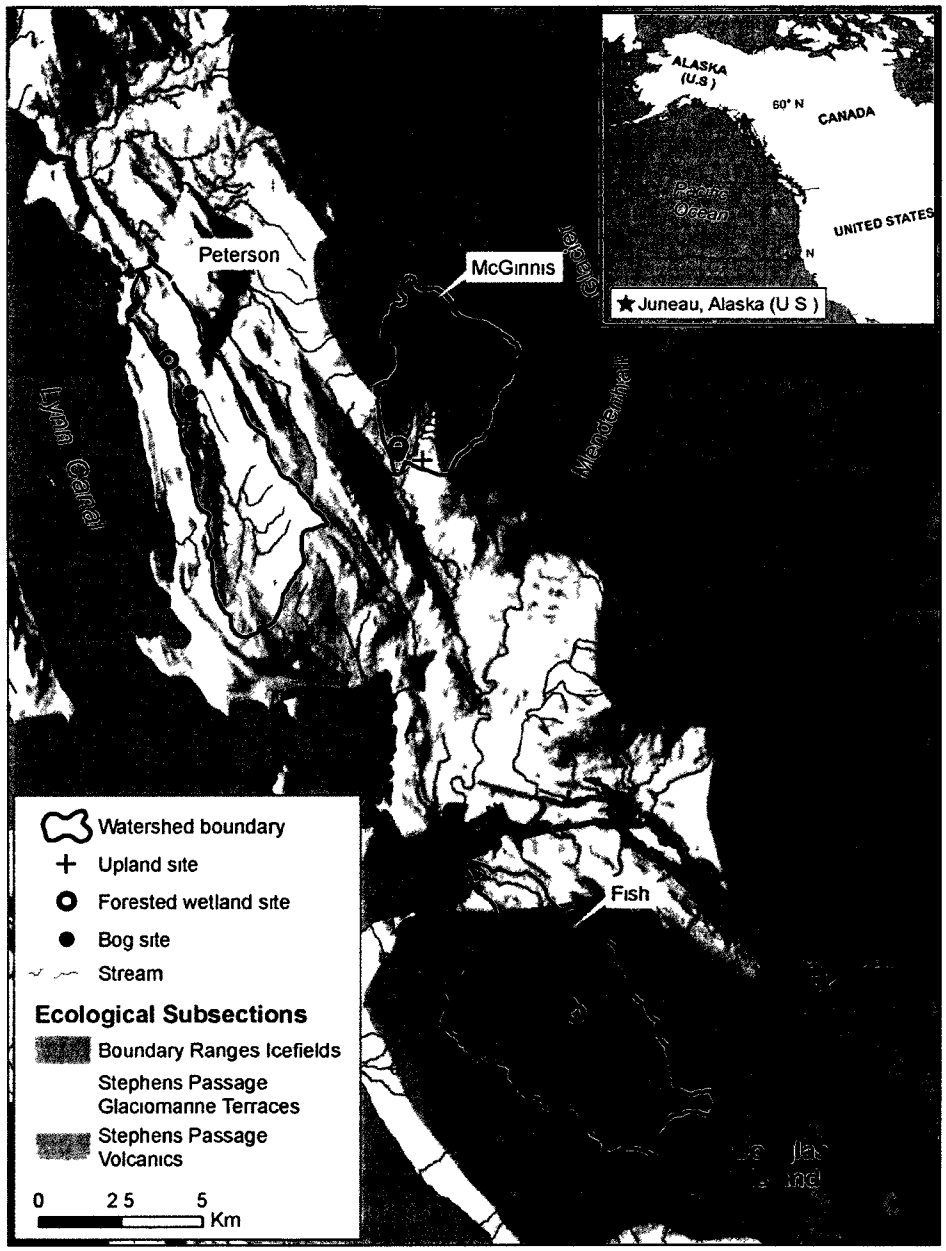

Figure 6.1. The ecological subsections and watersheds of the Juneau, AK area. The watershed boundaries define the approximate locations for estimates of net ecosystem production.

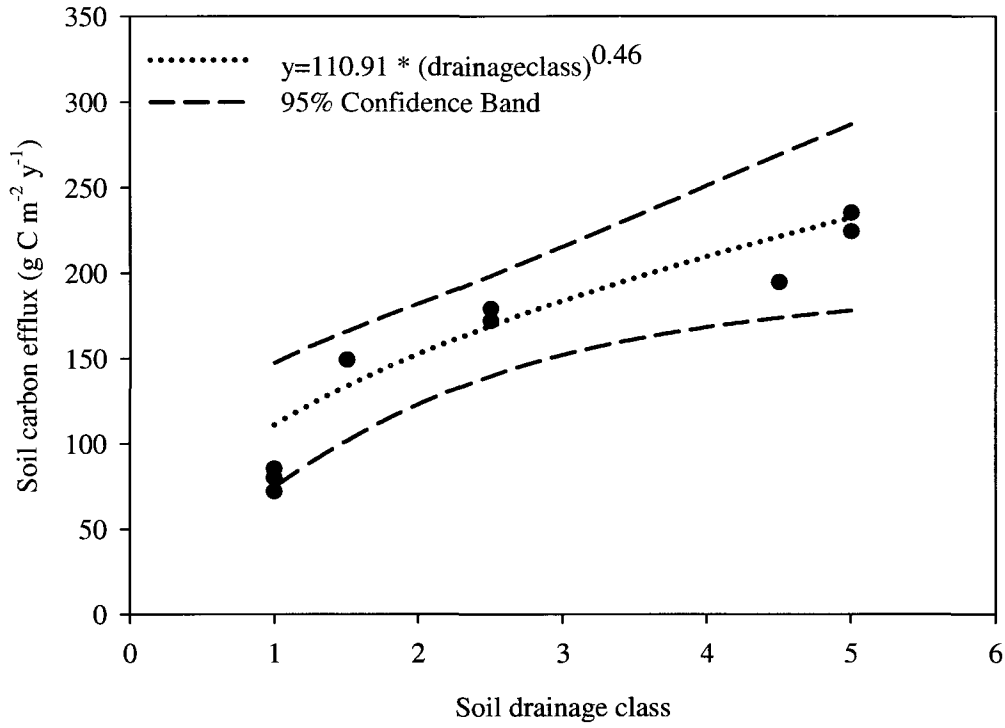

Figure 6.2. Predictive model for soil respiration efflux as a function of soil drainage class.

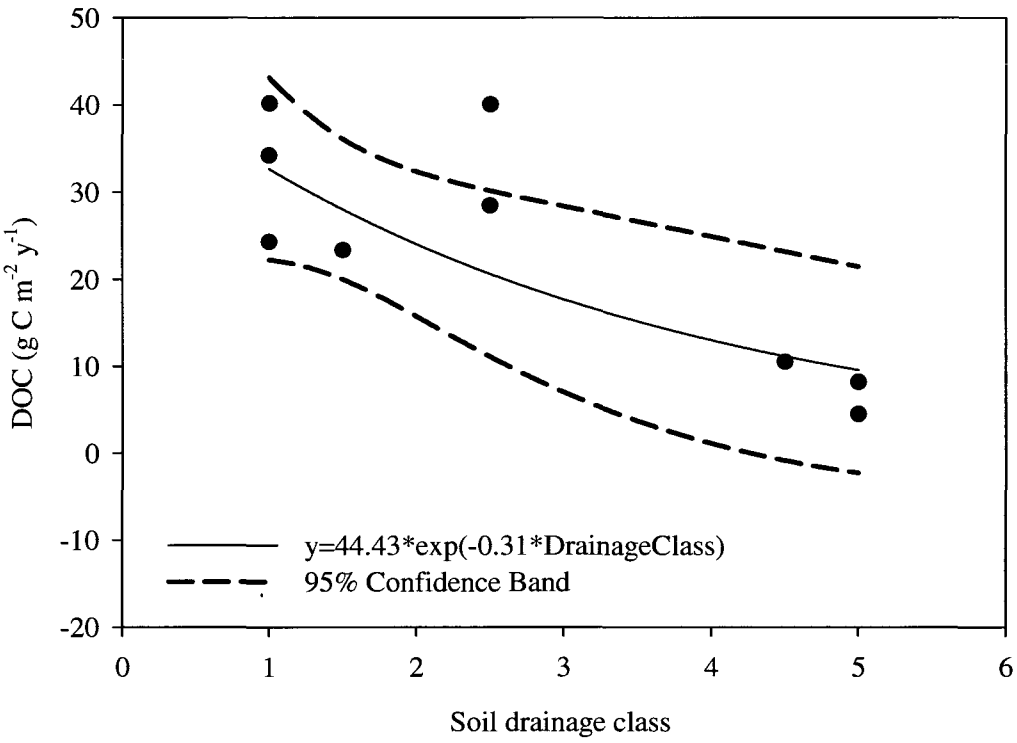

Figure 6.3. Predictive model for dissolved organic carbon (DOC) as a function of soil drainage class.

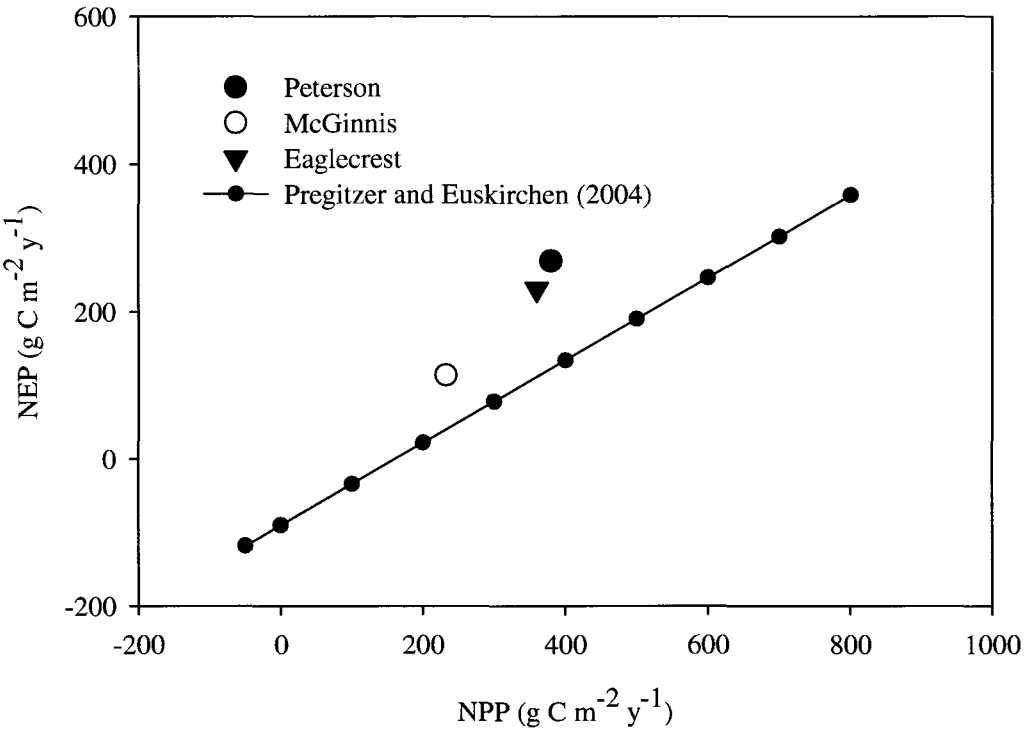

Figure 6.4. Relationship between NPP and NEP. Comparison of NPP:NEP in the current study with the predicted relationship generated from data for world biomes (Pregitzer and Euskirchen, 2004).

Table 6.1. Geomorphic subsection and maximum elevation of watersheds used to calculate net ecosystem production.

Watershed	Geomorphic subsection[†]	Elevation Max (m)
Eaglecrest	Stephens Passage volcanics	1563
McGinnis	Boundary Ranges Icefields	2236
Peterson	Stephens Passage glaciomarine terrace	931

[†]Nowacki et al., 2001

Table 6.2. Estimates of net primary productivity derived from MODIS[†] observations.

			Net Primary Productivity (g C m^2 yr^{-1})			
Watershed	Total area (ha)	Area measured (ha)	Min	Max	Mean	SD
Eaglecrest	3574	1862	163.1	511.7	388.80	80.77
McGinnis	1690	1690	27.8	423.9	259.75	117.73
Peterson	2474	2474	235.9	452.4	378.64	42.70

[†]Moderate resolution imaging spectroradiometer; NTSG (2011).

Table 6.3. Total carbon output calculated from soil respiration, DOC, POC, and DIC. Total carbon output from watersheds is based on estimates of carbon fluxes from hydropedologic types (smu) and drainage class (dc).

Watershed	Soil respiration				DOC		POC[§]		DIC		Total C output			
	Rt_{smu}†	Rt_{dc}‡	Rh_{smu}	Rh_{dc}	DOC_{smu}	DOC_{dc}	POC_{smu}	POC_{dc}	DIC_{smu}	DIC_{dc}	$Flux_{smu}$	$Flux_{dc}$	$Flux_{smu}$	$Flux_{dc}$
	Mg C ha^{-1}													
Eaglecrest	1.7	1.9	0.87	0.93	0.21	0.18	0.05	0.05	0.45	0.45	1.08	1.11	1.58	1.61
McGinnis	2.1	2.1	1.06	1.05	0.13	0.12	0.03	0.03	0.24	0.24	1.19	1.17	1.46	1.44
Peterson	1.3	1.3	0.65	0.67	0.29	0.29	0.07	0.07	0.09	0.09	0.94	0.96	1.10	1.12

†smu = hydropedologic type carbon flux based on drainage class assignment in three classes: sloping bog; forested wetland; upland.

‡dc = drainage class carbon flux based on predictive model of soil respiration or DOC as a function of drainage class.

§POC = particulate organic carbon, based on ratio of DOC:POC reported by Hope et al. (1994) from DOC values in hydropedologic type and drainage class calculations.

Table 6.4. Watershed carbon balance calculated from NPP and modeled total carbon output.

Watershed	NEP_{smu}	NEP_{dc}	NEP_{smu}†	NEP_{dc}†	Range‡
	----------Mg C ha^{-1}----------				
Eaglecrest	2.81	2.78	2.31	2.28	-0.17-4.04
McGinnis	1.40	1.42	1.13	1.15	-0.89-3.05
Peterson	2.85	2.83	2.69	2.67	1.42-3.58
Average ± SE	2.35 ± 0.82	2.34 ± 0.80	2.04 ± 0.81	2.03 ± 0.79	

†Estimate including POC and DIC.

‡Range based on carbon balance for minimum and maximum NPP without POC or DIC.

Chapter 7: Conclusions

The hydropedologic modeling approach developed in this dissertation provides a useful method to address the continuous nature of terrestrial ecosystem heterogeneity and associated function in the North American perhumid coastal temperate rainforest (NCTR). Carbon budgets in NCTR watersheds can be estimated by application of hydropedology to carbon cycling in soils. Carbon fluxes from soils are related to specific ecosystem types, and the strong hydrologic gradient in soils provides a basis for designing carbon budget models for NCTR watersheds. Seasonal fluctuations of soil water tables and temperature influence carbon storage and loss in soils of the NCTR. Annual carbon export of soil respiration efflux and dissolved organic carbon from watersheds can be estimated by combining soil saturation and temperature fluctuations. The combination of these major carbon export pathways and estimates of particulate and inorganic carbon flux constrain the vector of carbon losses from representative ecosystem types.

The conceptual model provided reasonable estimates of the carbon balance of three NCTR watersheds based on carbon fluxes from terrestrial ecosystem types based on hydrologic classes. The estimates of carbon accumulation establish evidence for a sink of carbon in mature forest and peatland ecosystem of the NCTR. Estimates of net ecosystem production ranged between 1.13 to 2.69 Mg C ha^{-1} among the watersheds with an average of 2.04 ± 0.81 Mg C ha^{-1}. The average estimate calculated from data in the dissertation compare quite well with both worldwide and regional values for forest carbon accumulation rates. The watershed estimates indicate that the terrestrial ecosystem is accreting carbon consistent with values for early (100-300 y) conifer forest development, yet lose a large portion of NPP through both gaseous and dissolved pathways.

The hydropedologic model established a means to capture and estimate all the forms of carbon loss from NCTR watersheds. The export of DOC was a key contributor to the total carbon budget estimate in watersheds in the NCTR. Thirty percent of the total carbon flux in wetland-dominated watersheds was via DOC. Hydropedologic modeling also provided a template for designing future experiments to test hypotheses about biogeochemical shifts under changing climates.